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Zoned olivines and their petrogenetic significance.[1]

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I. Introduction.

THE object of this paper is to demonstrate the widespread occurrence of zoned olivines in igneous rocks, to show that the zoning is continuous, that the composition varies from a magnesium-rich centre to an iron-enriched margin, and that these findings have an important bearing on the question of petrogenesis.

I propose to express the composition of olivine by stating only the percentage of fayalite, as the rock-forming olivines belong mainly to the forsterite-fayalite series. There are, however, a number of names in use for the members of this series. Assuming that the two names forsterite and fayalite are insufficient for the needs of classification, it yet seems to me that all but four of the existing names are superfluous. As the series is isomorphous and as the lines of division are purely arbitrary, I suggest the following decimal classification in molecular percentages:

Forsterite	0Fa— 10Fa
Chrysolite	10Fa— 50Fa
Hortonolite	50Fa— 90Fa
Fayalite	90Fa—100Fa

The name hyalosiderite was proposed in 1823 by Walcher for a ferruginous variety of chrysolite. At that time it was considered to be the

[1] Read at the Cambridge meeting of the British Association, August 1938, abstract in Rep. Brit. Assoc. Adv. Sci., 1938, p. 418–419.

most ferruginous variety of olivine. As fayalite was described and named in 1840 by Gmelin and hortonolite in 1869 by Brush, the name hyalosiderite is unnecessary and is besides inaccurate, since the mineral it attempts to describe is neither hyaline nor is it a siderite.

Although zoned olivines have been previously recorded and described their widespread occurrence has never been recognized, and, as a rule, they receive no mention in the text-books.

If we except a few records of zonally altered olivines by F. Zirkel (1), K. Hofmann (2), L. van Werveke (3), and J. J. H. Teall (5, p. 164), the first name demanding notice is that of B. Doss (4), who not only described zoned olivines from Syrian basalts and noted the variation of polarization-colours in the various zones, but also attempted to give an interpretation of this phenomenon as being due to variation in the chemical composition of olivines. His conclusions were that zoned olivines have an iron-rich centre and a magnesium-rich margin. In this he was supported by F. Möhle (10) and W. Schultz (11), who based their conclusions on the zonally disposed alteration products. It will be shown later on that olivines with a magnesium-rich margin are found in nature, but only as exceptions to the general rule.

The discovery of the relation existing between the optical properties and the chemical composition of olivines by S. L. Penfield and E. H. Forbes (6, 7) and the subsequent additional data provided by M. Stark (12), H. Backlund (14), and N. H. Magnusson (19) have provided mineralogists with a useful optical method for the study of olivines. In their final form the curves showing the relation between the optical properties and the chemical composition of forsterite-fayalite series are given by A. N. Winchell (31, p. 191).

A. Sigmund (8) was the first to apply the optical method to the study of zoned olivines from Styria. He found that (+) 2V of the outer zone is by 7° greater than that of the centre and that the outer zone shows higher polarization-colours than the centre. He therefore concluded that the outer zone is more ferruginous than the centre. F. Becke (9) also found that in the olivine from Predazzo melaphyre the difference of 2V between the centre and the margin was 7°, and in the olivine from the quartz-basalt of Cinder Cone, California, the difference was 5°, and that the outer zone in all cases was more ferruginous than the centre. M. Stark (13) gave a detailed account of zoned olivines from the basalts of the island of Ustica (NW. of Palermo, Sicily). In one of the crystals described by him the calculated amount of fayalite in the centre was $19\frac{1}{2}\%$ and in the margin $34\frac{1}{2}\%$. H. Michel (17) described some zoned

rtificial fayalites in which the centre was less ferruginous than the
argin. H. Backlund (18, 20) described a zoned olivine with 19Fa in
he centre and 36Fa in the margin from a nepheline-basalt (onkilonite)
om Vilkitzky Island, Arctic Ocean. Recently zoned olivines have been
ound to be of frequent occurrence in the Siberian traps (dolerites and
asalts) which probably belong to the Arctic igneous province (35, 40,
1). In a further description of zoned olivines, this time from the rocks
f Etna and Bohemia, M. Stark (23) made the first attempt to find a
elation between the composition of the olivine and the composition of
s parent rock. In this he was followed by F. Rinkel (25) and T. F. W.
arth (30). The last author also described a zoned olivine with 18Fa
the centre and 25Fa in the margin. Zoned olivines have also been
escribed by W. Campbell Smith and L. J. Chubb (26) from Austral
slands, Pacific Ocean, by P. Esenwein (28) from the Azores, by O. H.
dman (29) from Mt. Elgon, East Africa, by E. Jérémine (36) from
ake Rudolf, East Africa, by H. Kuno (38, 39) from Kurile Islands,
pan, and by L. Jugovics and A. Marchet (42) from Ságberg, Hungary.

A theoretical explanation of zoning in olivines was first given by C.
oelter who, in his 'Handbuch der Mineralchemie' (15), attributed this to
ie progressive enrichment of the growing crystals in a more fusible iron
thosilicate. J. H. L. Vogt (21) pursued this theory further and tried to
xplain zoning on the basis of a tentative thermal equilibrium diagram for
ie binary system forsterite-fayalite which he assigned to type I of
oozeboom, but it is only recently that N. L. Bowen and J. F. Schairer
7) have succeeded in establishing this diagram on experimental data.

Various Methods of Observing and Measuring Zoned Olivines.

Zonality in olivines can be ascertained by the following four methods:
) extinction-angles, (2) degree of birefringence, (3) indices of refraction,
) optic axial angles.

(1) The first method is very suitable for spotting zoned olivines in thin
ctions, but does not provide, as yet, an easy method to determine the
mposition of zonal components. Olivine, like all the orthorhombic
ystals, gives straight extinction (as measured from the pinakoidal
eavages) only in sections perpendicular to the principal planes of the
ptical ellipsoid. The angle of extinction in any other section can easily
determined graphically on a stereographic projection with the help
the law of Biot, which states that the vibration-direction in any
ction of a biaxial crystal is given by the bisectors of the angle between
e optic axes as projected on the plane of the section. To illustrate the

variation of the extinction-angle in a zoned olivine a theoretical case is given in which the centre has 2V = (+) 80° and the margin 2V = (−) 80°. Fig. 1*a* shows the curves of equal extinction measured from the trace of the dominant cleavage (010). The curves in fig. 1 *b* are

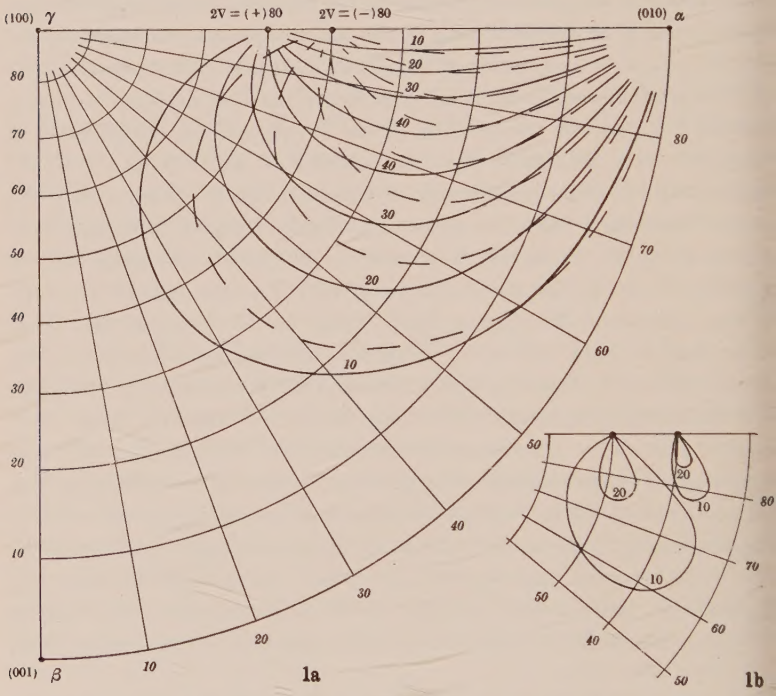

Fig. 1 *a*. Curves of equal extinction-angles from the cleavage (010) in an orthorhombic crystal for optic axial angles (2V) of (+) 80° and (−) 80°.

Fig. 1 *b*. Curves of equal differences of extinction-angles for same.

derived from the curves of equal extinction and represent the values of equal differences between the extinction of the centre and that of the margin. It is obvious, therefore, that we should find the greatest contrast of extinction-angles in the vicinity of the emergence of the optic axes, e.g. in the region of low birefringence. This is supported by actual observation, namely, that one can easily spot a zoned olivine in a thin section of rock by observing the variation of extinction-angle in grains showing grey polarization-colours. I found this method infallible, but it is quite obvious that when the difference in the composition of different zones is small, the movement of the extinction-shadow is slight

he determination of the optic axial angle from the angle of extinction theoretically possible if the correct orientation of the section is known. involves complicated calculations or the use of specially prepared iagrams, and so this method, as applied to the determination of the ptic axial angle, can hardly be advised.

(2) The variation in polarization-colours in different zones of markedly oned olivines is a very striking phenomenon, but only olivines having considerable zonal range show it. In the forsterite-fayalite series we et the following values for birefringence and polarization-colours in the rincipal planes of the optical ellipsoid in sections 0·02 mm. thick:

Plane.	Forsterite.	Fayalite.
$\perp\alpha$	0·020 Orange I	0·009 Blue-grey I
$\perp\beta$	0·035 Greenish-blue II	0·050 Red II
$\perp\gamma$	0·015 Yellow I	0·041 Yellowish-green II

In olivines with normal zoning, e.g. magnesia-rich centre and iron-ch margin, the colour will rise from the centre to the periphery in ctions perpendicular to β and γ, and fall in sections perpendicular to α. his means that in the majority of sections the colours will rise from the ntre to the periphery, but it is important to determine the orientation the section before drawing any conclusion. Great difference in colour n only be expected in the case of marked compositional difference tween the centre and the periphery. In the Irish Tertiary dolerites d basalts, for example, olivine varies on the average from 20Fa in the ntre to 40Fa at the margin. In this case the greatest variation in refringence and polarization-colour will be shown in sections perpen-cular to γ with $\beta-\alpha$ 0·019 in the centre and 0·023 at the margin. Thus sections 0·02 mm. thick the colour will vary from orange I in the ntre to orange-red I at the margin—a hardly perceptible variation. thicker sections naturally the variation of colour will be more striking, t, on the whole, olivines from Irish or Scottish Tertiary igneous rocks not show colour variation in a convincing way. On the other hand, ivines from certain Scottish essexites and from a luscladite from Rapa, ustral Islands, show a striking variation of colour. These rocks will discussed in the next part of this paper. One may conclude, there-re, that the observation of polarization-colours is only useful for spot-ng zoned olivines of a wide compositional range. The measurement of refringence in orientated sections may, however, provide a help in the termination of the composition.

(3) The refractive indices of the members of the forsterite-fayalite ries exhibit a continuous variation with the composition, rising with

the increase of iron in the mineral. The measurement of refractive indices in orientated sections of zoned olivines will no doubt give an approximate composition of the various zones, but such a procedure is cumbersome and time-robbing. An interesting effect has been observed by placing a drop of liquid of a suitable refractive index over an uncovered section of a zoned olivine fractured across the middle. On raising the objective the Becke line moved inwards into the mineral in the centre and outwards at the margins, and assumed the form of a sine curve.

(4) The optic axial angle, like the refractive indices, varies with the composition in the forsterite-fayalite series. Compared with the other methods, the determination of the optic axial angle presents the simplest and the surest means of arriving at the chemical composition of olivines. We do not know as yet how far the presence of other components affects the values of the optic axial angle, but their amount in the rock-forming olivines is very small and on the whole we can rely on the mean value curve given by A. N. Winchell (31, p. 191).

The apparent optic axial angle (2H) can easily be measured on the Fedorov universal stage. The connexion between the apparent (2H) and the true optic axial angle (2V) is given by the equation $\sin H = \beta \sin V/n$, in which β is the mean index of refraction of the mineral and n is the index of refraction of the glass hemispheres of the Fedorov stage. The obvious difficulty confronting us in this work is that of finding the mean index of refraction of the mineral actually measured on the Fedorov stage, especially when this mineral is a zoned one. To overcome this difficulty I have calculated the values of 2H for all the values of 2V and β as given on Winchell's diagram. Table I gives the results of this calculation made for Fedorov stage glass hemispheres of refractive index 1·649. This table enables us to read the value of 2V and the corresponding value of fayalite percentage from the measured value of 2H, and provides us with the basis for all subsequent calculations. One can also trace the curve of 2H above the curve of 2V on the diagram of Winchell and use the curve instead of the table.

The results obtained during the present work seem to confirm the data provided by Winchell. A number of olivine crystals whose optic axial angles were measured on the Fedorov stage were also tested for refractive indices and birefringences and no important discrepancy was noted, except in the case of fayalite, which will be discussed later. In my opinion, the optic axial angle gives the best index of the composition of olivine, and this method is far simpler and more accurate than any of the other methods.

The majority of olivine grains were measured from the emergence of oth optic axes, but when only one axis was available, an accurate etermination of an additional principal plane of the optical ellipsoid llowed this angle to be derived by construction on the stereographic rojection. The determination of the optic sign of the direction measured s essential in all this work.

TABLE I. Apparent ($2H\alpha$) and true ($2V\alpha$) optic axial angles for forsterite-ayalite series, together with the estimated percentages of fayalite, for Fedorov age hemispheres of refractive index 1·649.

		Fayalite %				Fayalite %	
$2H\alpha$.	$2V\alpha$.	Mol.	Weight.	$2H\alpha$.	$2V\alpha$.	Mol.	Weight.
94	94	0	0	73	66½	64	72
93	92	8	11	72	65½	66	74
92	90	13	18	71	64½	68	75
91	88½	17	23	70	63½	70	77
90	87½	20	27	69	62½	72	79
89	86½	23	30	68	61½	74	80
88	84½	27	35	67	60½	76	82
87	83	30	38	66	59½	78	84
86	82	33	42	65	58½	80	85
85	81	35	44	64	57½	81	86
84	79½	38	47	63	56½	83	87
83	78½	41	50	62	55½	85	89
82	77	44	53	61	54	87	91
81	75½	46	55	60	53	89	92
80	74½	48	57	59	52	91	94
79	73½	51	60	58	51	93	95
78	72½	53	62	57	50	95	96
77	71	56	65	56	49	96	97
76	70	58	67	55	48	98	98
75	68½	60	68	54	47	100	100
74	67½	62	70				

The method of determination of the apparent optic axial angle is as llows: the β-direction of the optical ellipsoid of the crystal is brought to coincidence with the K-axis of the stage, the stage placed in a agonal position and slowly rotated around the K-axis. The position of e optic axes in zoned olivines will be marked by shadows moving from e centre of the crystal to the periphery, or vice versa, according to the sition of α- and γ-axes. First, the maximum extinction of the central rtion of the crystal is noted and the angle (k) recorded, then the stage slowly rotated around the K-axis, the moving shadow fixed and the gle recorded for any desired position within the crystal. The difficulty fixing the maximum shadow on the actual margin of the grain is con-derable, and I believe that all the values in this paper relating to e composition of the margin are slightly underestimated. A further

complication is due to the fact that quite a number of olivine crystals have their outer zone replaced wholly or in part by alteration products (serpentine, iddingsite, chlorophaeite, &c.), and these obscure the vision for measurement purposes. In respect of the central parts of olivine crystals, the measurements again fall short of true values, but in a direction opposite to those of the margin. This is due to the fact that only a few sections pass through the true centre of the grain, but in this case the probable error is not very considerable. We may take it, therefore, that the measured compositional range of zoned olivines is somewhat below its true value.

III. Optical Data.

I have tried for the purpose of this section to determine within the compass of one single olivine crystal the difference in composition from zone to zone and to do this by the measurement of certain optical values along specific distances within the crystal. The olivines selected for these determinations are strongly zoned ones: (1) from luscladite, Rapa, Austral Islands, described by W. Campbell Smith and L. J. Chubb (26); and (2) from essexite, Crawfordjohn, Lanarkshire.

(1) Olivine from Rapa luscladite.—The measurement of the optic axial angle and the extinction-angle from the trace of (010) cleavage, at regular intervals from the centre to the margin, gave the following results:

	Distance in mm.	2H.	Angle of extinction.
Centre	0·0	90°	15°
	0·1	90	15½
	0·2	89	16
	0·3	87	17
	0·4	83	20
	0·5	77	24
Margin	0·6	70	29

The determination of birefringence in this section cannot be made with any degree of accuracy, but two orientated olivine grains from the same rock gave the following values:

	$\gamma-\alpha$.	$\gamma-\beta$.
Centre	0·039	0·016
Margin	0·049	0·012

(2) Olivine from Crawfordjohn essexite.—The most spectacular example of a zoned olivine was found in essexite, Crawfordjohn, Lanarkshire (fig. 2). Unfortunately, the section is mounted on a large-sized glass slide and thus not suitable for measurement on the Fedorov stage. A further difficulty is that the phenocryst of olivine is trisected by cracks,

but this defect has been remedied on the photograph. The pole of the section is inclined to the *c*-axis of the crystal at approximately 20°. The dominant cleavage (010) is parallel to the shorter diagonal of the rhomb, while the less prominent cleavage (100) is parallel to the longer diagonal.

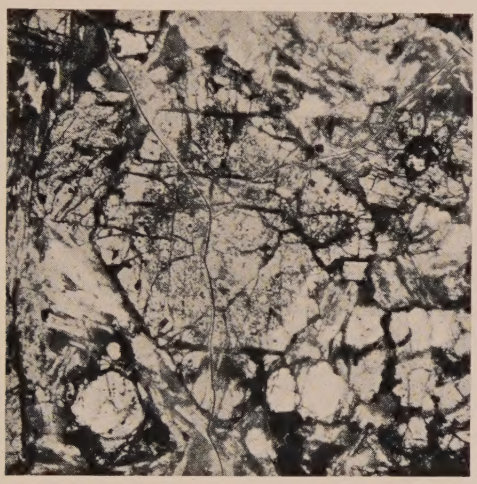

FIG. 2. Zoned olivine in essexite, Crawfordjohn, Lanarkshire. Ordinary light. Magnification 25.

The crystal faces cut by the section belong to the prism zone. They are (120) and (110), and their unequal development explains the asymmetry of the rhombic section. The inner portion of the crystal is full of inclusions of iron-ore and orthite, but the outer zone, some 0·2 mm. thick, is free from inclusions. The polarization-colour of the inner part is blue and that of the outer zone yellowish-green with a narrow bluish-green transition zone. The following values of the birefringence ($\gamma'-\alpha'$) were obtained by means of Berek's compensator:

	Distance in mm.	$\gamma'-\alpha'$.
Centre	0·0	0·027
	0·1	0·027
	0·2	0·027
	0·3	0·027
	0·4	0·028
	0·5	0·029
	0·6	0·030
	0·7	0·034
Margin	0·8	0·041

Thus the measurement of the optic axial angle, birefringence, and angle of extinction along measured distances in zoned olivines shows that

the variation of these values, and therefore the variation in composition, is roughly proportional to the cube of the distance from the centre of the crystal. When plotted on diagrams (not given in this paper) with x-axis corresponding to the distance from the centre and y-axis to the difference of 2H, angle of extinction, or birefringence, the curves obtained show a good approximation to the cubic curve $y = x^3$. This means that the change in composition in successive zones is proportional to the volumes occupied by these zones, in other words, the change of composition per unit volume is a constant value in a given olivine crystal. The average composition of a zoned olivine, therefore, will be equal to the arithmetic mean of the composition of the central part and that of the margin.

The birefringence was also measured in some other sections, mainly for the purpose of checking the validity of Winchell's curves. A section of olivine from Crawfordjohn essexite perpendicular to α gave: centre $\gamma-\beta$ 0·017, 2H 86½°; margin $\gamma-\beta$ 0·016, 2H 81½°. While another perpendicular to β gave: centre $\gamma-\alpha$ 0·037, margin $\gamma-\alpha$ 0·041.

The birefringence $\gamma-\alpha$ was also measured for the olivine from the three varieties of dolerite of Fair Head, Co. Antrim, which will be discussed later. The values obtained were as follows:

	Centre.	Margin.
Olivine from glomeroporphyritic dolerite	0·034	0·038
,, ,, olivine-dolerite	0·038	0·042
,, ,, dolerite-pegmatite	0·043	0·047

All these results seem to show that we get a fairly good agreement between the values for birefringence and the values for the optical axial angle as given in Winchell's diagram.

The values for the optic axial angle were determined for 88 olivine crystals from various igneous rocks (table II). For convenience, the results are translated into fayalite molecular percentages, but any one interested in the actual angles can easily obtain them by referring to table I. The 88 olivines given in table II are further arranged in 13 groups according to the nature, age, and locality of the rocks, and the averages of these groups are given in table III, together with the difference between the composition of the centre and that of the margin, and also the average of the two values.

The olivines measured were not chosen on account of their zonality, but for their orientation in sections suitable for measurement on the Fedorov stage. Considering the variability of composition shown by olivines, even those belonging to one and the same rock, it may be

TABLE II. Composition of zoned olivines calculated from the optic axial ngle.

roup.	No.	Name of rock and locality.	Fayalite Mol. % Centre.	Margin.
I	1.	Peridotite inclusion in basalt, Calton Hill, Derby-	8	8
	2.	shire.	13	13
II	3.	Olivine-dolerite, Wooden Hill, Jedburgh, Roxburgh-	33	34
	4.	shire.	35	44
	5.		44	53
	6.	Olivine-dolerite, Peniel Heugh, Jedburgh, Rox- burghshire.	20	30
	7.	Olivine-dolerite, Rutherford, Kelso, Roxburghshire.	20	23
	8.	Olivine-dolerite, Duns, Berwickshire.	33	34
	9.		20	21
	10.	Olivine-dolerite, Cottonshope, Northumberland.	20	35
	11.		38	48
	12.		38	48
	13.		35	53
	14.		44	51
III	15.	Olivine-basalt (Dunsapie type), Queenscairn, Stich- ill, Roxburghshire.	0	17
	16.	Olivine-basalt (Dunsapie type), Lurdenlaw, Kelso, Roxburghshire.	38	53
	17.	Olivine-basalt (Dunsapie type), East Softlaw, Kelso.	53	54
	18.		27	30
	19.		48	50
IV	20.	Teschenite, Spalefield, Anstruther, Fife.	13	20
	21.		13	20
	22.		15	28
	23.		18	25
	24.		23	38
	25.	Teschenite, Gosford Bay, East Lothian.	13	33
	26.		20	38
	27.		20	44
	28.		27	48
	29.		41	58
V	30.	Essexite, Crawfordjohn, Abington station, Lanark- shire.	13	38
	31.		32	45
	32.		33	58
	33.		38	52
	34.		38	53
	35.		38	53
	36.		44	58
VI	37.	Glomeroporphyritic dolerite, Fair Head sill, Bally- castle, Co. Antrim.	0	20
	38.		8	35
	39.		20	20
	40.		23	35
	41.		27	27
	42.		27	56

Group.	No.	Name of rock and locality.	Fayalite Mol. % Centre.	Margin.
VII	43.	Olivine-dolerite, Fair Head sill.	27	44
	44.		30	56
	45.		38	66
VIII	46.	Dolerite-pegmatite, Fair Head sill.	62	74
	47.		64	80
	48.		64	83
IX	49.	Olivine-dolerite, Portrush sill, Co. Antrim.	8	38
	50.		20	33
	51.	Olivine-dolerite, Falbane dike, Ballycastle, Co. Antrim.	13	27
	52.		13	33
	53.	Olivine-dolerite, North Star dike, Ballycastle, Co. Antrim.	13	44
	54.		20	38
	55.		20	53
	56.		27	44
	57.		33	53
X	58.	Olivine-basalt (lower plateau lava), The Gobbins, Island Magee, Co. Antrim.	13	33
	59.		13	41
	60.		17	30
	61.		20	33
	62.	Olivine-basalt (lower plateau lava), Clogh, Ballymena, Co. Antrim.	0	23
	63.		35	46
XI	64.	Olivine-basalt, Madras, Oregon, U.S.A.	33	44
	65.		33	53
	66.		35	51
	67.	Olivine-basalt, Pocatello, Idaho, U.S.A.	13	38
	68.		27	34
	69.		30	41
	70.	Olivine-quartz-basalt, Cinder Cone, California, U.S.A.	18	42
	71.		23	29
	72.		23	41
	73.		27	33
	74.		35	46
XII	75.	Luscladite, Tapui Islet, Ahurei Bay, Rapa, Austral Islands, Pacific Ocean.	20	70
	76.		33	81
	77.		33	85
	78.		44	87
	79.		44	89
	80.		48	85
	81.		48	87
	82.		48	89
	83.		53	89
	84.		58	96
	85.		58	96
XIII	86.	Fayalite-quartz-porphyry, Kudaru Hills, Nigeria.	100	100
	87.		100	100
	88.		100(?)	100(?)

objected that the sampling was insufficient and that more measurements should have been made. Although it is true that a larger number of

neasurements would increase the accuracy of the results, the time taken y these measurements would prove excessive (the measurement and lotting of each grain takes from one to two hours) while the additional ata would probably alter the average but slightly. In order to test this made 25 additional measurements of the olivine from the Cottonshope livine-dolerite. The result of these additional measurements shows a otal range of 23Fa to 53Fa and an average range of 33Fa to 46Fa, while he average of the tabulated data (table II, nos. 10–14) gives a total range f 20Fa to 53Fa and an average range of 35Fa to 47Fa. The difference is egligible if we take into account the probable error of measurement.

ABLE III. Calculated average composition of zoned olivines in different rock-types.

	Group.	Rock-type.	Mol. % of fayalite. Centre.	Margin.	Difference.	Average.
ritish arboni- rous- ermian cks	I	Peridotite	10½	10½	—	10½
	II	Olivine-dolerite	31½	39½	8	35½
	III	Olivine-basalt	33	41	8	37
	IV	Teschenite	20	35	15	27½
	V	Essexite	33½	51	17½	42
ish ertiary cks	VI	Glomeroporphyritic dolerite, Fair Head	17½	32	14½	25
	VII	Olivine-dolerite, Fair Head	31½	55½	24	43½
	VIII	Dolerite-pegmatite, Fair Head	63½	79	15½	71
	IX	Olivine-dolerite	18½	40½	22	29½
	X	Olivine-basalt	16½	34½	18	25½
	XI	Olivine-basalt, Western States, U.S.A.	27	41	14	34
	XII	Luscladite, Rapa, Austral Islands	44½	87	42½	65½
	XIII	Fayalite-quartz-porphyry, Kudaru Hills, Nigeria	100	100	—	100

Although individual olivines found in the same or similar rocks ndoubtedly vary in composition, this variability has a definite range. n the other hand, in olivines drawn from different rock-types the nge and average composition differ from type to type. Compare, or example, the three phases of the Fair Head sill (table III, verages VI, VII, and VIII) for dissimilar ranges; and averages II, II and IX, X, XI for similar rock-types. Certainly one must realize at the measured olivines are either phenocrysts or larger grains the groundmass, and that small grains and ophitic olivines of late rystallization could not be measured. The probable effect of this lection is to lower the range on the fayalite side. Ophitic olivines nclosing felspar laths are extremely common among the Irish Tertiary

basalts and dolerites and they are invariably richer in iron than the idiomorphic phenocrysts. On the other hand, an analysis of olivine extracted from a rock shows a higher percentage of magnesia than an analysis of the same olivine calculated from the analysis of the rock. This may be partly due to the removal of the highly ferruginous iddingsite borders during the process of extraction of the mineral from the rock, as I have already suggested (34, p. 506), but it may be also due to the removal of the outer ferruginous border and also to the selection of larger crystals for analysis.

The examination of tables II and III shows that the rocks selected vary from ultrabasic to acid. The ultrabasic end is represented only by a peridotite inclusion from a basalt. The olivine in this inclusion is very rich in magnesia. This agrees with a previously published analysis of this olivine (27, p. 713) which, on recalculation, gives 14Fa. Examination of other peridotites confirms the impression that, on the whole, the olivine in them is rich in magnesia, but apparently there are peridotites, like those of the Bushveld complex in South Africa, in which the olivine is more ferruginous. In dolerites and basalts the average content of fayalite is much higher and varies between 25 and 40. In more alkalic rocks, such as essexite and luscladite, the fayalite content is still higher and the zonality more marked. Unfortunately, in typical teschenites the olivine is too decomposed for the purpose of optical determination. The teschenites given in table II have more affinities with the crinanites, and their olivines are comparable with those of dolerites and basalts. Olivines from Carboniferous dolerites and basalts are not strongly zoned, although these rocks are more alkaline than those from the Irish Tertiary in which good zoning is common. Columbia River plateau basalts, as well as others from the Western States of America, are very similar to the Irish Tertiary basalts and possess well-zoned olivines. In the Siberian basalts and dolerites the olivine is conspicuously zoned (35, 40, 41), and it would be interesting to find if the olivines from the North Atlantic region are zoned and to what degree. The greatest range of zoning is shown by the olivine from luscladite of Rapa, Austral Islands. This olivine is also ferruginous (table III, average XII). Three olivine grains from a fayalite-quartz-porphyry of Kudaru Hills, Nigeria, described by A. D. N. Bain (33) correspond to pure fayalite. Two of them gave 2H 54°, a figure which, according to table I, corresponds to 100Fa; but one grain gave 2H 50°. The same low angle has been obtained in a fayalite from a slag. This may be due to the presence of some other constituents or to the imperfection of the olivine diagram.

The composition of zoned olivines reduced to six averages is repre-ented graphically in fig. 3. The range of zoned olivine in each average s represented by an arrow drawn from the point corresponding to the omposition of the centre to that corresponding to the composition of

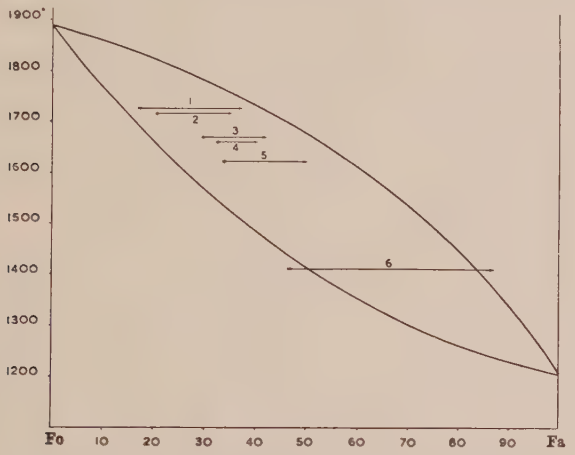

Fig. 3. Average compositional ranges of zoned olivines from different regions plotted against the thermal equilibrium diagram for the binary system forsterite-fayalite. 1, Irish Tertiary dolerites and basalts; 2, British Carboniferous-Permian teschenites; 3, Basalts of the Western States of America; 4, British Carboniferous-Permian dolerites and basalts; 5, British Carboniferous-Permian essexite; 6, Luscladite, Rapa, Austral Islands.

ıe margin. These arrows are plotted against the thermal diagram for .e forsterite-fayalite series as given by N. L. Bowen and J. F. Schairer 7). With the exception of the Rapa luscladite, all the average ranges zoned olivines fall within the boundaries of the liquidus and solidus ırves. That the zonal ranges do not stretch right across the theoretical nge marked by the curves is what one would expect, considering the obability of error in measurement and the possibilities of incomplete uilibria and resorption of crystals formed at an earlier stage of crystal-ation. In spite of these shortcomings, the present investigation fully pports the experimental results obtained by Bowen and Schairer and ovides an illustration of the crystallization process as outlined by them. ıe crystallization of olivines in igneous rocks on the whole follows e course from forsterite to fayalite, and both the individual crystals d the successive crops of crystals tend to become more ferruginous

as the crystallization proceeds. This trend may be called the normal course of crystallization and the consequent zoning—the normal zoning. There are, however, cases of oscillatory and reverse zoning, in which the normal zoning is followed by a reversal to an early magnesium-rich phase. Only a few cases of reverse zoning were noted during the present work, and unfortunately the crystals were not suitable for measurement. It is interesting to note, however, that in the case of reverse zoning the outermost magnesium-rich zone is sharply separated from the inner core. Reverse zoning was the one which was observed by B. Doss (4), F. Möhle (10), and W. Schultz (11), who concluded that it represented the common type of zoning. It was left to A. Sigmund (8) to demonstrate the significance of the normal zoning.

IV. Petrogenetic Significance of Zoned Olivines.

Rock-forming olivines belong mainly to the forsterite-fayalite series. Statistical study shows that out of 238 analysed olivines from igneous rocks and meteorites 195, or 82%, contain more than 90% of Mg_2SiO_4+Fe_2SiO_4. It is only olivines from slags that are rich in Ca_2SiO_4 and Mn_2SiO_4. A frequency-distribution curve constructed on the basis of 257 analyses of olivines (including 19 analyses of olivines from slags) is given in fig. 4. In my opinion this frequency-distribution curve does not correspond to the real frequency-distribution of olivines in nature. This assumption is based on the following argument. A rough estimate indicates that basic rocks with an average olivine content of 15% form about one-third of the total of the igneous rocks of the earth's crust; and although olivine-rich ultrabasic rocks are extremely rich in olivine their relative abundance is negligible as compared with the basic rocks, being about 0·5% of the total of the igneous rocks. The amount of olivine present in the intermediate and acid rocks is far too small to be of any account. Therefore the mode of the frequency-distribution curve should fall at the average composition of olivines from the basic rocks. The average composition of olivine from dolerites and basalts as computed from their optical data (table III) is about 30Fa (mol.) or 38Fa (weight). On the other hand, the mode of the frequency curve (fig. 4) corresponds to 15Fa (weight). In my opinion this mode is lowered through the unconscious selection of olivines analysed, the majority of which are olivines from peridotites and phenocrysts from basalts, both of them exceptionally rich in magnesia. Unconscious selection is probably likewise responsible for the rise of the curve at the fayalite end, as mineralogists are somewhat attracted by rare minerals which they are eager to

nalyse on every possible occasion. Actually, fayalite is found only casionally and in minute quantities in alkalic intermediate and acid cks. It is the character of the distribution of analyses that makes . H. Magnusson (19) suggest an immiscibility gap between hortonolite

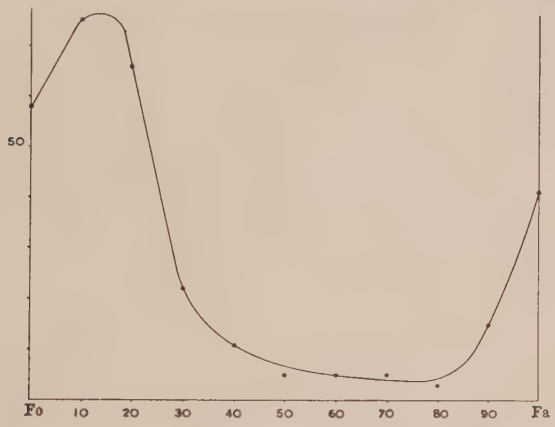

FIG. 4. Frequency-distribution curve for 257 analyses of olivines recalculated into weight percentages of forsterite and fayalite.

d fayalite. Neither the results of the work of Bowen and Schairer) nor the data provided in the present paper tend to support this ggestion. In the zoned olivine from the Rapa luscladite (table II, s. 75–85) the composition runs from chrysolite to hortonolite (20Fa–Fa) and from hortonolite to fayalite (58Fa–96Fa). Olivine from the ir Head dolerite-pegmatite (table II, nos. 46–48) also partly covers e supposed immiscibility gap. The apparent scarcity of olivines in the erval between Mg-hortonolite and fayalite may be due to the difficulty procuring suitable specimens.

There have been several attempts to correlate the composition of vines with the composition of the containing rocks. B. Doss (4), who lieved that olivines tend to develop a magnesium-rich outer zone, plained this phenomenon as being due to an early crystallization of n-ore and a consequent enrichment of the residual magma in magnesia. the other hand, M. Stark (23) arrived at the conclusion, which is w shared by many petrologists, that the residual magma is pro-ssively enriched in FeO at the expense of MgO, and that the olivines ow a parallel tendency. Stark has demonstrated that olivines in lim-rgites are richer in magnesia than olivines in basalts, and that olivines

in acid rocks are very rich in iron (23, p. 38). F. Rinkel (25) has also concluded that olivines from peridotites are the richest in magnesia and those from eulysite are the richest in iron, while those from basalts and gabbros occupy an intermediate position. The origin of eulysite is still a debatable point, but according to the original interpretation by H. von Eckermann (22) eulysite represents a pegmatitic residual solution of basic magma, i.e. a late crystallization product. In this connexion it is interesting to inquire into the nature and composition of fayalite-bearing rocks. L. Hawkes (24) has fully discussed this subject and published a table of 10 analyses of these rocks. To these I have added 12 recent analyses of fayalite-bearing granite, granophyre, porphyry, and phonolitic trachyte. The average and the range of composition of these 22 analyses gave:

SiO_2	65·20%	(56·59–75·55)
Na_2O	5·42	(3·21– 7·59)
K_2O	4·58	(1·59– 6·63)

$FeO(Fe_2O_3) : FeO(Fe_2O_3) + MgO = 0·95 : 1.$

This shows that fayalite-bearing rocks are alkalic and range from intermediate to acid, and, in the majority of cases, are quartz-bearing. There are, however, certain basic rocks in which olivine and quartz are co-present (olivine-bearing quartz-dolerites), but olivine in these rocks, as a rule, is only moderately rich in iron, although in some varieties, according to V. Sobolev (32), it may contain as much as 80Fa. The average of three analyses of olivine-bearing quartz-dolerite gives: SiO_2 52·10, Na_2O 2·54, K_2O 1·19%, Fe/Mg ratio 0·73; and in this respect they do not differ from normal quartz-dolerites. According to Bowen and Schairer, fayalite can exist in equilibrium both with quartz and albite, as is actually the case in fayalite-bearing rocks; but in the olivine-bearing quartz-dolerites, olivine is too magnesian to be in equilibrium with quartz and its presence is explained better if we assume that it represents an untransformed relict of an early phase of crystallization. Therefore, one may agree with Bowen and Schairer when they say that the fayalite-bearing rocks 'are always rich . . . in alkali feldspars alone or together with quartz or nepheline and thus belong among the trachytes, rhyolites, and phonolites' (43, p. 408). Recently, A. B. Edwards (44, p. 299), from his measurements of the optic axial angle of olivines from a suite of rocks ranging from limburgite to trachyte, also comes to the conclusion that olivine becomes more fayalitic as the magma approaches the trachytic end of the suite.

To find the relation between the composition of the olivine and the

mposition of the parent rock I have calculated the Fe/Mg ratio for me of the rocks studied in this paper and the corresponding ratios of livines as derived from their optical data. The results of these calcula-ons are given in table IV and fig. 5.

TABLE IV. Calculated ratio $FeO(Fe_2O_3) : FeO(Fe_2O_3)+MgO$ for rocks and ivines.

	No.	Rock-type.	Rock.	Olivine.
	1.	Glomeroporphyritic aggregates (= allivalite, Rum), Fair Head	0·27	0·37
ish	2.	Olivine-dolerite, Fair Head	0·56	0·57
ertiary	3.	Dolerite-pegmatite, Fair Head	0·65	0·82
cks	4.	Olivine-dolerite	0·57	0·38
	5.	Olivine-basalt	0·57	0·42
ritish				
arboni-	6.	Teschenite	0·64	0·40
rous-	7.	Olivine-dolerite	0·60	0·49
ermian	8.	Essexite	0·57	0·56
cks				
	9.	Luscladite, Rapa	0·46	0·76
	10.	Fayalite-bearing rocks	0·95	0·98

The data for the Rapa luscladite have been obtained from the analysis ublished by W. Campbell Smith and L. J. Chubb (26), that for the air Head dolerite and dolerite-pegmatite from unpublished analyses by e author. Unfortunately, there is no analysis as yet of the glomero-rphyritic aggregates from the Fair Head sill and so an analysis of llivalite from the Isle of Rum was taken instead, as the rocks are very milar in composition. The remaining ratios were derived from the lculated averages of the appropriate analyses.

The diagram (fig. 5) provides us with a clear picture of the relation tween olivines and rocks in respect of their relative proportions of eO and MgO. In the majority of dolerites and basalts the olivines are cher in MgO than their parent rocks. Considering that a fairly large roportion of iron is removed from the magma in the form of ore this what one would expect. Also from the nature of the thermal equi-brium diagram one would expect the early formed olivine to be richer magnesia. The three points corresponding to the three phases of the air Head sill clearly show this effect. They also show the progressive nrichment in iron of the olivines in the successive cooling phases, rresponding to the enrichment in iron (relative to magnesia) in the ck as a whole.

T. F. W. Barth (30, p. 380) has given a similar diagram in which the nount of olivine present in the rocks is also shown. His conclusions

are that in the olivine-rich rocks olivine is more magnesian than in the olivine-poor rocks, and that the olivines of the phenocrysts are always richer in magnesia than those of the groundmass. The material given in the present paper is in complete agreement with the conclusions of Barth.

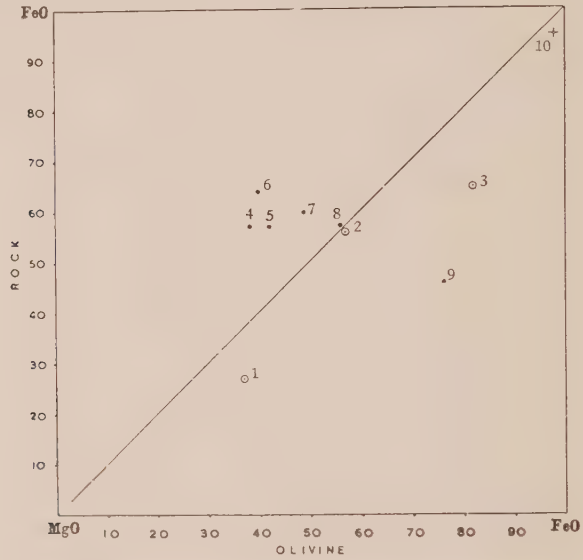

FIG. 5. Diagram showing the relation between the ratio $FeO(Fe_2O_3) : FeO(Fe_2O_3)+MgO$ calculated for the olivines and their parent rocks.

1. Glomeroporphyritic aggregates (= allivalite) } Fair Head sill } Irish Tertiary rocks.
2. Olivine-dolerite
3. Dolerite-pegmatite
4. Olivine-dolerite
5. Olivine-basalt
6. Teschenite } British Carboniferous-Permian rocks.
7. Olivine-dolerite
8. Essexite
9. Luscladite, Rapa, Austral Islands.
10. Fayalite-bearing rocks.

In this respect olivines are similar to pyroxenes, the composition of which, according to A. Weich (16), also depends on the composition of the rock in basic rocks pyroxenes are relatively rich in magnesia and get progressively richer in iron as the acidity of the rock increases.

All this shows that there exists an intimate connexion between the composition of the magma and the composition of olivines crystallizing

ut of it at different stages of solidification or differentiation. There is lso a parallelism between the compositional change in individual rystals of olivine as shown by zoning and that of the successive crops f olivine crystals in a cooling magma. Without inquiring into the actual rocess of differentiation of magma, which may be different in different ases, one must stress the importance of the control exercised by the iagma on the minerals and the reciprocal control exercised by the inerals on the magma.

In conclusion, I wish to thank Lt.-Col. W. Campbell Smith, Dr. C. A. nderson, Dr. A. D. N. Bain, Dr. C. E. Marshall, and Dr. A. C. Waters r sending me rock-specimens.

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Merosymmetry versus merohedrism.

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[Read November 3, 1938.]

WHILE crystal classes based upon symmetry furnish a clear-cut classification of crystals free from ambiguity and uncertainty, there are cogent reasons for recognizing the larger groups called systems.

It is important to recognize crystal systems, not only on account of physical properties, but also on account of the difficulty of assigning some crystals to their appropriate class. For example, a crystal consisting of a tetragonal prism and a pinakoid may belong to any one of seven classes. For the time being it is placed in the tetragonal system. When it is carefully investigated and its symmetry determined it may be assigned to the proper class.

Contrary to statements that have been made,[1] we can have, and do have, a two-fold classification in geometrical crystallography—the thirty-two classes based upon symmetry, and the six systems based upon properties that certain classes have in common.

In the opinion of the writer the most satisfactory set of names for the crystal classes is the one based upon names of general forms. These names of Groth are comparatively simple, consistent, and lend themselves to international usage. Moreover, they are entirely free from ambiguity and this can scarcely be claimed for any other set of names. The use of these class names in Groth's 'Chemische Krystallographie' (1906–19) indicates that in all probability they will become standard, since this important five-volume reference work is so extensively used by structural crystallographers the world over.

Class names based upon merohedrism or merohedry (general terms to include both hemihedrism and tetartohedrism or the corresponding terms hemihedry and tetartohedry) have been very widely used for the past century, but many crystallographers have abandoned these terms since they lead to inconsistencies.[2]

[1] Victor Goldschmidt (trans. by M. A. Peacock), Amer. Min., 1931, vol. 16, pp. 18–33. [M.A. **4**–439.]

[2] W. J. Lewis A treatise on crystallography. Cambridge, 1899, p. 259.

Application of the principle of merohedrism failed to establish all of possible crystal classes. At one time, for example, sucrose was igned to a possible tetartohedral division of the orthorhombic tem.[1] Hemimorphism was formerly placed on an entirely different ting from other kinds of hemihedrism.[2] In spite of the preliminary earches of Hessel and Bravais, it was not until the work of the Finnish stallographer Gadolin in 1871 that we had a really satisfactory ssification of crystals. By the use of the single principle of symmetry dolin deduced the thirty-two classes of symmetry.

And not until the symmetry classes were derived was it possible to ablish the crystal systems satisfactorily. The diclinic system of the idle nineteenth century disappeared, not because of the impossibility diclinic axes of reference (I have recently suggested that the term linic syngony be used for such cases[3]), but because there were no stal classes that could be assigned to it. Gadolin used seven systems, Fedorov, by means of his syngony-ellipsoid, showed that there are y six crystal systems, or syngonies as he preferred to call them. orov's work has been confirmed by the writer, who defines the six stal systems in terms of zones.[4]

Although often discarded and regarded as having historic interest y, terms involving merohedrism sometimes have been retained for lagogic reasons. The late A. H. Phillips,[5] for example, says: 'The old nod of deriving the hemihedrons, etc., from the holohedrons has n retained and may meet with criticism, but it is a simple method letermining what forms are possible to combine on crystals of lower metry. I have always found it a very material aid to the student, ving no false impressions.' Now, one's recognition of merohedrism ns to depend largely upon the emphasis placed upon crystal systems the question arises as to whether crystal systems or crystal classes more fundamental. If space permitted, arguments might be given how that crystal classes are more fundamental than crystal systems. mere fact that there is disagreement as to the number of crystal tems is almost proof of this.

But even if this is admitted, there is need of a series of general terms classes of varying degree of symmetry within the crystal systems.

[1] G. H. Williams, Elements of crystallography. New York, 1890, p. 169.
[2] P. Groth, Physikalische Krystallographie. First edition, Leipzig, 1876.
[3] A. F. Rogers, Amer. Min., 1935, vol. 20, p. 210.
Idem, Introduction to the study of minerals. Third edition, New York, 1937, 38–39.
[4] A. H. Phillips, Mineralogy. New York, 1912, p. v.

This need has usually been supplied by the use of terms based upo merohedrism or merohedry. The class with the highest grade of sym metry in a system was designated holohedral, those obtained by sup pressing one-half of the faces and extending the other half were calle hemihedral, and those obtained by extending one-fourth of the face were called tetartohedral.

Since the selection of faces is always a symmetrical selection, and thi is rarely stated, much less emphasized, *it seems advisable to use the term merosymmetry*, *hemisymmetry*, and *tetartosymmetry*, rather than mero hedrism, hemihedrism, and tetartohedrism. The key-note of geometrica crystallography, if I understand it aright, is symmetry. The first chapte of Fedorov's Курсъ кристаллографіи (Course of crystallography, 3r edition, St. Petersburg, 1901) is entitled О симметріи (On symmetry

The term *merosymmetry* (from the Greek μέρος, part or fraction) wa introduced by Nevil Story-Maskelyne in a series of lectures before th Chemical Society of London, which appeared in the Chemical News[1] fo 1875. These lectures formed the basis for Story-Maskelyne's classi work 'Crystallography, a treatise on the morphology of crystals' whic appeared in 1895. Since the publication of this important book the ter holosymmetry often has been used, due largely to the influence of Miers and Spencer,[3] but to the best of my knowledge there has been no particu lar recognition of the value of Story-Maskelyne's idea. This is probabl due largely to the fact that Story-Maskelyne himself was not altogethe consistent. Instead of using hemisymmetric, tetartosymmetric, an ogdosymmetric (he used six crystal systems, so there is one case o ogdosymmetry) he introduced such terms as haplohedral and diplohedra holosystematic and hemisystematic, which unnecessarily complicate his nomenclature. However, Story-Maskelyne did clearly set forth satis factory divisions of the crystal systems and was one of the first to recog nize the importance of the thirty-two symmetry classes.

The purpose of this communication is to advocate the general use o the terms holosymmetry and merosymmetry with its subdivision hemisymmetry, tetartosymmetry, and ogdosymmetry. By substitutin the suffixes *-symmetry* and *-symmetric* for *-hedrism* and *-hedral* we hav a satisfactory series of terms. We then have the advantages claimed fo merohedrism by Phillips and at the same time emphasize symmetry. I is a pity that this suggestion was not made a quarter of a century ago

[1] N. Story-Maskelyne, Chem. News, London, 1875, vol. 31, pp. 3–4 . . . 232.
[2] H. A. Miers, Mineralogy. London, 1902.
[3] L. J. Spencer, Encyclopaedia Britannica, 11th edition, 1910, vol. 7, p. 569.

t geometrical crystallography will doubtless be studied for centuries come. If the proposal is a good one, the change should be made even ough it is rather late.

On the table of the thirty-two crystal classes (p. 256) I have attempted put down what I consider to be a simple consistent set of terms based on merosymmetry. In the left-hand column are the six crystal stems with the two sub-systems of the hexagonal system. In the next umn are the class numbers which I have used for some time.[1] These slight modifications of Groth's numbers[2] arranged so as to place sses 21 and 22 in the hexagonal sub-system instead of in a separate gonal system as Groth did. The class numbers are convenient for erence and it should be emphasized that the sequence of the classes only in minor part arbitrary. The arrangement is such that every ss has a lower class number than any of the classes of which it is a sub-up.[3]

Next are the class names of Groth which are based upon Fedorov's nes of general forms. The names of the isometric classes are simpli-; hexoctahedral, for example, instead of hexakisoctahedral. In the rth column are what Shearer, the structural crystallographer, has pily called the symmetry number, which will be recognized as the nber of faces in the general form or what in group theory is called the er of the group (point-group).

Next are the class names based upon merosymmetry. Since there are crystal systems there are six holosymmetric classes. Asymmetric sed for the pedial class instead of hemisymmetric, because strictly aking there is no symmetry at all, hence no hemisymmetry. In the noclinic system the adjectival terms axial and planar are adapted n Boldyrev.[4] The convenient term holoaxial was introduced by rs.[5] Polar is used, but not equatorial, since it implies the use of the eographic projection. The terminology should, I believe, be inde-dent of any particular method of projection. The term alternating has been used, since it seemed to be unnecessary. For two of the classes, . 19 and 23, triple names unfortunately seemed to be necessary. While the terms I have used here are probably free from ambiguity,

[1] A. F. Rogers, Amer. Min., 1928, vol. 13, p. 573. [M.A. 4–54.]
[2] P. Groth, Physikalische Krystallographie. 4th edition, Leipzig, 1905.
[3] A. F. Rogers, Proc. Amer. Acad. Arts Sci., 1926, vol. 61, pp. 198–200. [M.A. 39.]
[4] A. K. Boldyrev and V. V. Dolivo-Dobrovolsky, Ann. Inst. Mines, Leningrad, , vol. 8, p. 155. [M.A. 6–79.]
[5] H. A. Miers, Mineralogy. London, 1902, p. 45.

Table of the thirty-two Crystal Classes

System.	Class. no.	Class names (based upon the general form).	Sym. no.	Alternative names (based upon merosymmetry).	Example.
Triclinic	1	Pedial	1	Asymmetric	$H_2Sr(C_4H_4O_6)_2.4H_2O$
	2	Pinakoidal	2	Holosymmetric	Albite
Monoclinic	3	Sphenoidal	2	Axial hemisymmetric	$C_{12}H_{22}O_{11}$(Sucrose)
	4	Domatic	2	Planar hemisymmetric	Clinohedrite
	5	Prismatic	4	Holosymmetric	Gypsum
Orthorhombic	6	Rhombic disphenoidal	4	Holoaxial hemisymmetric	Epsomite
	7	Rhombic pyramidal	4	Polar hemisymmetric	Hemimorphite
	8	Rhombic dipyramidal	8	Holosymmetric	Barite
Tetragonal	9	Tetragonal disphenoidal	4	Sphenoidal tetartosymmetric	Cahnite
	10	Tetragonal pyramidal	4	Polar tetartosymmetric	$Ba(SbO)_2(C_4H_4O_6)_2.H_2O$
	11	Tetragonal scalenohedral	8	Sphenoidal hemisymmetric	Chalcopyrite
	12	Tetragonal trapezohedral	8	Holoaxial hemisymmetric	$NH_3(CH_3)I$
	13	Tetragonal dipyramidal	8	Dipyramidal hemisymmetric	Scheelite
	14	Ditetragonal pyramidal	8	Polar hemisymmetric	$AgF.H_2O$
	15	Ditetragonal dipyramidal	16	Holosymmetric	Zircon
Hexagonal — Rhombohedral sub-system	16	Trigonal pyramidal	3	Ogdosymmetric	$NaIO_4.3H_2O$
	17	Rhombohedral	6	Rhombohedral tetartosymmetric	Dioptase
	18	Trigonal trapezohedral	6	Holoaxial tetartosymmetric	α-Quartz
	19	Ditrigonal pyramidal	6	Polar ditrigonal tetartosymmetric	Tourmaline
	20	Hexagonal scalenohedral	12	Rhombohedral hemisymmetric	Calcite
Hexagonal — Hexagonal sub-system	21	Trigonal dipyramidal	6	Trigonal tetartosymmetric	—
	22	Ditrigonal dipyramidal	12	Trigonal hemisymmetric	Benitoite
	23	Hexagonal pyramidal	6	Polar hexagonal tetartosymmetric	Nepheline
	24	Hexagonal trapezohedral	12	Holoaxial hemisymmetric	β-Quartz
	25	Hexagonal dipyramidal	12	Dipyramidal hemisymmetric	Apatite
	26	Dihexagonal pyramidal	12	Polar hemisymmetric	Iodyrite
	27	Dihexagonal dipyramidal	24	Holosymmetric	Beryl
Isometric	28	Tetartoidal	12	Tetartosymmetric	Ullmannite
	29	Gyroidal	24	Holoaxial hemisymmetric	—
	30	Diploidal	24	Central hemisymmetric	Pyrite
	31	Hextetrahedral	24	Polar hemisymmetric	Tetrahedrite
	32	Hexoctahedral	48	Holosymmetric	Garnet

realize that other terms could have been chosen. There are in some ses a number of choices. For example, class 3 might have been called loaxial hemisymmetric or polar hemisymmetric; class 13 might have en designated as central hemisymmetric or equatorial hemisymmetric. is difficulty in naming a class will serve to emphasize the great value Groth's names of classes which are names of general forms. In consequence of this only one name for each class is possible after the form mes are established. You will note that I regard names based upon erosymmetry as alternative names which are often useful, especially discussing the crystal classes of a particular system. The classes are the time being regarded as divisions of crystal systems. But these mes are not to replace class names based upon general forms.

Then finally, in the last column, examples are given of substances rstallizing in each class, except for classes 21 and 29. Acid silver osphate is sometimes given for class 21, but there has never been any oof that it belongs in this class. Cuprite and sylvite have been assigned class 29, but they almost certainly are holosymmetric. Other candics for class 29 are γ-Al_2O_3, and γ-Fe_2O_3, but Dr. Tom Barth[1] is ost certain that they are holosymmetric, since they are probably morphous with each other, and γ-Al_2O_3 forms solid solutions with losymmetric $MgAl_2O_4$. Cahnite, $Ca_4B_2As_2O_{12}.4H_2O$, from Franklin rnace, New Jersey, belongs to the tetragonal disphenoidal class[2] and ce in the table it may replace $Ca_2Al_2SiO_7$, a laboratory product igned to this class by Weyberg.

We now have four useful sets of terms for the 32 symmetry classes:

1) Class names based upon general forms.
2) Alternative names based upon merosymmetry.
3) Example names of substances crystallizing in each class.
4) Class numbers.

ese are listed in order of their probable importance. We need onyms here as in other fields. Examples of mineral names were used Miers, who spoke of the calcite class, for example. As suggested lier in this article, I find the class numbers very convenient. After ne use they become quite familiar.

The French school of crystallographers following Bravais make the symtry of the space-lattice the basis of the crystal system and so employ en systems instead of six. This appeals to many modern structural stallographers. In this usage holohedral crystals are defined as those

[1] Personal communication, October 1938.
[2] Lazard Cahn, personal communication, May 1938.

having the same symmetry as the lattice and merohedral crystals as those with less symmetry than that of the lattice. In the opinion of some crystallographers this is the main justification for the use of merohedrism.

Now it is true that there are *seven* styles of crystal architecture and this should be emphasized even more than the 14 Bravais space-lattices. But to insist upon seven crystal systems based upon the symmetry of lattices is to give the term crystal system a meaning different from the generally accepted use of that term. In whatever terms it may be defined, a crystal system is thought of as a series of closely related crystal classes with geometrical and physical properties in common. The difficulty to which I refer does not seem to be appreciated by structural crystallographers. Let me explain it.

In five crystal systems (triclinic, monoclinic, orthorhombic, tetragonal, and isometric) there is no difficulty, for there is but one primary lattice for each system. But for five of the twelve classes with a single hexagonal zone there is a serious objection to this method of defining a crystal system. In the case of each of five classes of what I call the rhombohedral sub-system (nos. 16–20 of the tabulation on p. 256) there are two possible space-lattices, the hexagonal and the rhombohedral. Now this places us in the awkward position of knowing the crystal class of a given crystal before we know its crystal system. I contend that the determination of the symmetry of a crystal should automatically determine the crystal system. We cannot properly speak of systems of symmetry, but the symmetry classes are the constituent members of the crystal systems. One of the principal values of crystal systems is their use as groups larger than crystal classes, as suggested in the second paragraph of this article. It is important, I believe, to retain the original and generally accepted use of the term crystal system.

My friend and former student Dr. J. D. H. Donnay has kindly called my attention to a difficulty of another kind that is not so easy to solve this also concerns the rhombohedral sub-system. Calcite, for example is hemisymmetric according to my use of the term, since it belongs to the hexagonal system, but from the standpoint of the French school of crystallographers and many structural crystallographers it is considered to be holohedral, since the space-lattice is rhombohedral. This introduces a complication which I think can be cleared up by suitable nomenclature, but I have no suggestion to offer at present.

It is a pleasure to record my indebtedness to Dr. J. D. H. Donnay for a critical reading and discussion of the manuscript of this article.

Prismatic cleavage and steep rhombohedral form in α-quartz.

By Julien Drugman, M.Sc., PhD.

[Read November 3, 1938.]

1) An example of α-quartz showing good prismatic cleavages.

As is well known, quartz rarely shows well-defined cleavages. Kenngott has shown that quartz that has suddenly been cooled, er strong heating, has a tendency to break up in the directions of mbohedron faces. Breithaupt (cf. Hintze, Min., 1904, vol. 1, p. 1273) ntions 'cleavage following the primary rhombohedron and the prism, ely well defined, often passing into conchoidal fracture'.

. 1. Polished slice of α-quartz cut perpendicular to the *c*-axis, showing prismatic cleavages.

t may be useful to put on record an example of clear rock-crystal which cleavage is developed in all three 60° directions of the prism es. The specimen was originally in the Daimeries collection, in ssels, having been bought years ago at Idar, near Oberstein on the ne, by A. Daimeries. I obtained it from Mr. A. G. F. Gregory, who bought the Daimeries collection. The locality from which it came unfortunately, unknown.

The specimen (fig. 1) is a slice, 2 cm. thick, cut normally to the *c*-axis ough a large crystal of clear rock-crystal. It measures 17 cm. along

the longest diagonal with a width of 9 cm. The cleavage cracks continue right through the thickness of the slice and extend, in one case for a length of over 7 cm., in others for several centimetres in parallel orientation to one or other of the three prism faces. In some there is a tendency for the ends of the cracks to curve round, in two cases at least resuming a straight course following another of the three 60° directions. These must, then, be true cleavages, otherwise there would not be this tendency to resume a direction parallel to another prism face. This cleavage is, however, not an easy one.

Unfortunately nothing is known of the cause that produced the cleavages in this specimen. They may have been already present in the specimen when it was found or may have been developed by some intense mechanical action during the process of cutting and polishing. The surfaces of the cleavages reflect light strongly when placed in a suitable position, and they are seen to be fairly flat and smooth. One or two show a 'hackly' surface, being formed of flat bands elongated horizontally, and all show striations in the same horizontal direction. The flatter parts of these hackly fractures are joined by irregular curved parts. Neither in these nor in any of the other fractures in the slice is any orientation along a rhombohedron face discernible. Cleavage is observed in the prism directions alone, and none in the direction of the primitive rhombohedron. Other cracks show the usual curved, somewhat conchoidal forms that often occur in quartz.

Beside these cleavages and fractures in the thickness of the slice there are surface fractures that also show definite orientations, in each case along the main diagonal of the slice and parallel to one prism face. The most distinct are a set of deep troughs with curved sides, but all elongated in this same direction; then a series of close-packed surface-grooves continuing the troughs on one face of the slice. Finally, two well-defined conchoidal fractures can be observed on the main prism edge, each caused by a sharp blow on the edge.

A photograph (fig. 1) of the slice, somewhat reduced in size, shows the cleavages and surface fractures fairly satisfactorily, though an examination of the specimen itself can alone bring out all the details.

The absence of information as to the origin of these cleavages makes it difficult to decide whether these are really true cleavages or whether they may be in relation to twinning. Twinning would, I imagine, be more likely to develop rhombohedral parting, and no trace of this appears on the specimen. I would therefore regard it as being definitely in the nature of true cleavage.

) An example of α-quartz crystals with a steep rhombohedron as the predominant form.

The list of known crystal-forms on α-quartz is a long one. Yet the ain habit of this mineral is, as every one knows, an extremely constant ıe. There are few occurrences in which the association of the prism th the two unit rhombohedra is not the dominant one, the positive ombohedron being sometimes more important than the negative one. ıe other forms occur as modifications on the edges and corners of this mbination. Sometimes one or other of these other forms become more ominent and we obtain, for instance, quartz crystals with acute ramidal terminations through the predominance of a steeper rhombo- dron. A. Lacroix[1] figures crystals from Bourg-d'Oisans, Isère, in ıich the acute rhombohedron $(0.11.\overline{11}.1)$ is sometimes so important to become the predominant form. But other crystals of the same urrence show intermediate stages, and the acute rhombohedron can- t be considered the essential one of the combination.

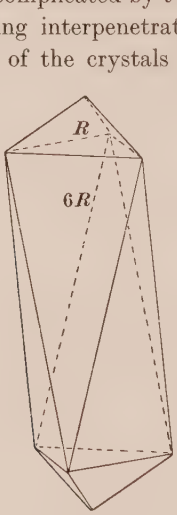
C. Palache[2] has described an occurrence, however, in which an acute ombohedron is present as an independent form, only occasionally odified by the two unit rhombohedra. These were found in a cavity the Franklin Furnace (New Jersey) ore, embedded in a felted mass he finest crocidolite. They ranged in size from minute sharp spicules rods 3 cm. long and 3 mm. in diameter, some with needle-sharp ngular terminations, others blunted by faces of the two unit rhombo- ira. The faces not being very brilliant, measurements were not ficiently precise to determine whether the acute rhombohedron was $(0.18.\overline{18}.1)$ or $(0.17.\overline{17}.1)$, and no sure means were available to ascer- n which of the two unit rhombohedra was the positive one. Palache ws, in his drawing, the first $(0.18.\overline{18}.1)$ of these steep rhombohedra. the larger development of the unit rhombohedron, that becomes the ative one on this assumption, suggests that, perhaps, the steep mbohedron $(0.17.\overline{17}.1)$ may be the true one. So far as I can discover, occurrence is the only one recorded in which a steep rhombohedron resent as a really independent or predominant form.

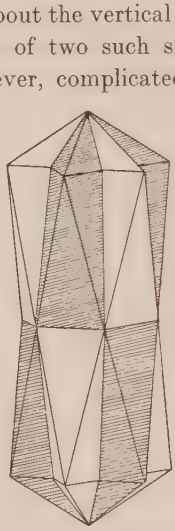
nother such case has come to my notice while examining, with a ocular microscope, minute quartz crystals occurring on a specimen tibnite from Kisbánya, Romania, that I had obtained from Mr. E. P. tley last June. The fine, bladed crystals of stibnite occur in closely ked sheaths on crusts formed of minute rhombohedra of chalybite

[1] A. Lacroix, Min. France, 1901, vol. 3, p. 86, fig. 56; p. 81, figs. 51 and 52.
[2] C. Palache, Amer. Min., 1928, vol. 13, p. 319, pl. VIII, fig. 23.

associated with more obtuse rhombohedra of calcite. One side of these crusts is in many places coated with a layer of very minute, closely intergrown, doubly terminated quartz crystals. These quartzes all show the same habit, and that a very unusual one, as we have just seen, of an acute rhombohedron modified by one very subordinate unit rhombohedron, which we may assume to be the primitive $+R\,(10\bar{1}1)$. The steep rhombohedron is of the same sign as the unit one and would, on the above assumption, be also positive.

Some of these crystals are of ideal symmetry (fig. 2). Most of them are, however, complicated by twinning about the vertical axis, the most symmetric being interpenetration twins of two such simple crystals (fig. 3). Most of the crystals are, however, complicated by frequent





α-Quartz, combination of R and $6R$, from Kisbánya, Romania.

Fig. 2. Simple, untwinned crystal.

Fig. 3. Interpenetration twin about the vertical axis.

repetitions of the same law of twinning and the juxtaposition of the numerous elements in parallel position to each other, producing the appearance of true sixfold symmetry. The edge between the two rhombohedra is more or less rounded, but no other form is present than these two. The acute rhombohedron is here, as in the occurrence described by Palache, a truly independent and predominant form.

The surface of the steep rhombohedron is somewhat rough and, in spite of good reflections from the primitive rhombohedron, no accurate readings could be obtained on the goniometer. Rough measurements on the microscope stage of the angle between adjacent faces, together

th the measurement of the acute angle at the apex of each rhombo-
dron face and of the ratio of the side to the short diagonal, give
fficient ground for taking the indices of this face to be either $(60\bar{6}1)$
$(70\bar{7}1)$. They agree slightly better with the first of these, and the
awings have in consequence been based on a rhombohedron $6R$ $(60\bar{6}1)$.
ie following table gives the results of these rough measurements:

	Measured.	Calculated value for $6R$.	Calculated value for $7R$.
gle between adjacent faces	58°±2°	59° 29′	59° 19′
ute angle of the rhombohedron face	24°±2°	25° 20′	22° 0′
tio of side to short diagonal	2–2½	2·28	2·62

The symmetrical development of these crystals points to their having
en formed in suspension in a homogeneous solution. They seem to
ve settled from this on to the carbonate crusts, occurring on these as
irregular mass of crystals, some of which are easily detached without
nage. They are, however, too minute to be mounted conveniently
the goniometer. They form very suitable material for study with the
ocular microscope. This is, in fact, a good example of the use that
n be made of 'micro-mounts' in the study of minute crystals that
uld otherwise pass unnoticed. For the complex intergrowths pro-
ced by repeated twinning are not easily elucidated, even by careful
amination with an ordinary single-tube microscope.

The crystallography of sartorite.

By F. A. Bannister, M.A., Adolf Pabst, Ph.D., and George Vaux, B.Sc., D.I.C.

[Read November 3, 1938.]

SARTORITE is one of four minerals once thought to be isomorphous. Of these only two are truly so—chalcostibite, copper antimony sulphide, $CuSbS_2$; and emplectite, copper bismuth sulphide, $CuBiS_2$—both orthorhombic minerals. The third, zinckenite, has been shown by two of the authors[1] to be hexagonal and to approach more closely the formula $Pb_6Sb_{14}S_{27} = 6PbS.7Sb_2S_3$ than $PbSb_2S_4 = PbS.Sb_2S_3$. It seemed appropriate to continue a study of the series by an X-ray investigation of sartorite, a lead arsenic sulphide formerly supposed isomorphous with zinckenite. More than usual interest was attached to such a task, since discrepant interpretations of the symmetry and setting of sartorite had been obtained by well-known crystallographers.

Indeed, the difficulties of indexing different crystals of sartorite on the same lattice proved insuperable to G. F. H. Smith and R. H. Solly,[2] so that they devised three incongruent lattices—one monoclinic and two anorthic. Also they used slightly different elements for each crystal. Their device, however, was not uniformly successful with certain crystals of sartorite previously measured by C. O. Trechmann.[3] Two crystals in particular eluded their efforts. These, they decided, could not belong to the same species, but must be another mineral, perhaps a dimorphous form of sartorite, which they then named sartorite-α.

Amongst an extremely fine collection of sartorite crystals from the Lengenbach quarry, Binn valley, Switzerland, preserved in the Department of Mineralogy of the British Museum, are fortunately the actual crystals measured by Trechmann, and by Smith and Solly. The first problem was to determine whether they would all give the same diffraction patterns. Several crystals were selected including Trechmann's nos. 1 and 2, renamed sartorite-α by Smith and Solly, and Trechmann's no. 5 on which Smith recorded over one hundred faces.

X-ray rotation photographs about the elongation axis [010] of all

[1] G. Vaux and F. A. Bannister, Min. Mag., 1938, vol. 25, p. 221.

[2] G. F. H. Smith and R. H. Solly, Min. Mag., 1919, vol. 18, p. 259.

[3] C. O. Trechmann, Min. Mag., 1907, vol. 14, p. 212.

ese crystals are completely superposable. The layer-lines although not ose together are packed with spots. It was considered necessary, there-re, to show that photographs of the same crystals rotating about the 00] or [001] axes are also superposable. This was only achieved by tting the crystals as accurately as possible with Lauegrams. These vealed orthorhombic symmetry, but owing to the high absorptive wer of lead minerals for X-rays the spots probably correspond only the pseudo-cell to which we shall refer shortly. The rotation photo-aphs of the crystals studied proved them all to be identical and esumably sartorite. No small differences in spacing could be detected m one crystal to another and certainly nothing approaching the riability of crystal elements suggested by Smith and Solly. Moreover, X-ray evidence exists that sartorite-α is an independent species. In respects Trechmann's crystals nos. 1 and 2 are identical with the rest.

Long-exposure oscillation photographs of a sartorite crystal about the es [100], [010], and [001] revealed a number of intermediate weak er-lines not detected on the rotation photographs. These are restricted the [100] and [001] axes, and they have different intensities above d below the equator, so that sartorite must really be monoclinic. No empt has been made to work out the monoclinic space-group, but ly the space-group of the well-marked pseudo-cell, the diffractions m which are due largely to the lead atoms. The weak spots, which only revealed by oscillation photographs, are, on the other hand, e entirely to diffractions from the much lighter arsenic and sulphur oms. The cell dimensions are a 58·38, b 7·79, c 83·30 Å. with β 90°. e orthorhombic pseudo-cell has the dimensions a 19·46, b 7·79, 4·17 Å. Space-group $P2_12_12_1$.

It should be stated here that no X-ray evidence of twinning in sartor-has been found. As for zinckenite, the crystals photographed give er-line patterns for all three axes in which: (a) all the layer-lines are actly parallel to the equator; and (b) the layer-lines of each photograph rrespond to integral sub-multiples of only one of the three cell sides b, or c. Twinning of the type found by Buerger[1] in mispickel cannot detected by X-ray methods, but should be observable on polished tions in polarized light. We have examined bright striated faces of ith and Solly's sartorite crystal no. 1, but could find no trace of twin-nellae nor any marked anisotropy.

A thorough optical investigation of other sartorite crystals has not en carried out, so that our evidence does not contradict in any way

[1] M. J. Buerger, Zeits. Krist., 1936, vol. 95, p. 83. [M.A. 6–409.]

H. Schneiderhöhn and Ramdohr's[1] account of twin-lamellae in the mineral. The optical and X-ray examination of crystal no. 1 show that this crystal at least is not twinned, and that the symmetry and cell-dimensions are most probably the true ones for untwinned sartorite.

Chemical analyses of sartorite, mostly prior to 1855, do not show any appreciable variations from the simple formula $PbAs_2S_4$, and the contents of the pseudo-cell are $3{\cdot}95PbAs_2S_4$. Rounding this figure off to 4, the true unit cell contents are then $240PbAs_2S_4$.

The relation of sartorite to the three minerals zinckenite, chalcostibite, and emplectite is not one of isomorphism, but the elongation direction in all four minerals is close to, or a simple multiple of, the corresponding spacing 3·89 Å. for stibnite. It is hoped to undertake a crystal-structure analysis of the pseudo-cells of zinckenite and sartorite, so that the interrelations of the four minerals may be revealed in more detail. A corresponding study of jordanite[2] and dufrenoysite, two lead arsenic sulphides with which sartorite was confused by the early crystallographers, is only one of the numerous corollaries. The few crystal-structures that have so far been worked out, together with our cell measurements, suggest that in the metallic arsenic, antimony, and bismuth sulphide minerals we have a group of substances rivalling in complexity and probably as closely interrelated as the silicates.

The final portion of this study is devoted to the reconciliation of axial ratios given by former workers with the X-ray cell dimensions of sartorite and to the successful indexing of the numerous forms previously observed.

Sartorite has been assigned no less than four different sets of axial elements. G. vom Rath[3] and Victor Goldschmidt[4] considered it to be orthorhombic. Trechmann gave it monoclinic elements; and Smith and Solly actually suggested that the forms be referred to three incongruent lattices, one monoclinic and two anorthic. Table I shows that all of these elements (excluding the anorthic elements of Smith and Solly) can be reconciled with the elements that correspond to the unit cell or to the pseudo-cell.

It is desirable that the crystallographic observations on sartorite be explained by reference to the structural elements. To this end the

[1] H. Schneiderhöhn and P. Ramdohr, Lehrbuch der Erzmikroskopie, 1931, vol. 2 p. 491. [M.A. **4**–434.]

[2] The cell dimensions of jordanite have recently been given by W. E. Richmond in a paper by C. Palache, Amer. Min., 1938, vol. 23, p. 829.

[3] G. vom Rath, Ann. Phys. Chem. (Poggendorff), 1864, vol. 122, p. 373.

[4] V. Goldschmidt, Index der Krystallformen der Mineralien, 1888, vol. 3, p. 320 Krystallographische Winkeltabellen, 1897, p. 320.

easurements of Smith and Solly's crystal no. 1 were recalculated and was found that nearly all faces can be given simple indices, provided at some are referred to the elements of the true cell and some to the ements of the pseudo-cell. The agreement of observed and calculated gles is no worse than that obtained by Smith and Solly using three congruent lattices. Moreover, large discrepancies or relatively high dices are found only with faces which are described by Smith and lly as 'bad' or 'striated'.

TABLE I. Comparison of axial ratios of sartorite.

	$a:b:c$.	Transformation.	$a:b:c$.
	True cell.		
nnister (1938) ...	58·38 7·79 83·30 Å.	—	7·49:1:10·69
m Rath (1864) ...	0·539:1:0·619	$12c:b:20a$	7·43:1:10·78
ldschmidt (1888) ...	0·771:1:0·956	$6c:a:\frac{1}{1}\frac{0}{2}\frac{0}{0}b$	7·44:1:10·82
,, (1897) ...	0·956:1:0·771	$6a:c:\frac{1}{1}\frac{0}{2}\frac{0}{0}b$	7·44:1:10·82
echmann (1907) ...	1·276:1:1·195 $\beta = 77° 48'$	$6a \sin\beta:b:9c \sin\beta$	7·48:1:10·52
ith and Solly (1919)	2·574:1:2·712 $\beta = 74° 38'$	$3a \sin\beta:b:4c \sin\beta$	7·45:1:10·46
	Pseudo-cell.		
nnister (1938) ...	19·46 7·79 4·17Å.	—	2·498:1:0·535
m Rath (1864) ...	0·539:1:0·619	$4c:b:a$	2·476:1:0·539
ldschmidt (1888) ...	0·771:1:0·956	$2c:a:\frac{5}{1}\frac{}{2}b$	2·482:1:0·541
,, (1897) ...	0·956:1:0·771	$2a:c:\frac{5}{1}\frac{}{2}b$	2·482:1:0·541
echmann (1907) ...	1·276:1:1·195 $\beta = 77° 48'$	$2a \sin\beta:b:\frac{9}{2}\frac{}{0}c \sin\beta$	2·493:1:0·526
ith and Solly (1919)	2·574:1:2·712 $\beta = 74° 38'$	$a \sin\beta:b:\frac{1}{5}c \sin\beta$	2·485:1:0·523

Table II shows the results of the recalculation for the [010] zone of ith and Solly's crystal no. 1. In this table only the simpler of the o new indices have been given in those cases where a choice was easy. is brings out a simple relation showing the influence of the pseudo- l on the morphology. In the zone [010] both axes of the pseudo- l differ in magnitude from those of the true cell. For faces close to 00) simpler indices are found when they are referred to the true cell ments, whereas for faces closer to (001) simpler indices are found en they are referred to the elements of the pseudo-cell.

The relations are somewhat different and possibly more striking in the 01] zone, table III. Here the ratio $a:b$ is much smaller, i.e. more arly 'normal' for the pseudo-cell, and all faces close to (100) receive mple indices only when referred to the elements of the pseudo-cell. If e elements of the true cell were used alone, no faces making an angle

of less than 39° 47′ with (100) could receive indices with h less than 10; in other words, faces with indices in this range could only occur in a little more than half of the zone.

TABLE II. Indexing of the faces in the [010] zone of sartorite.

$(h0l)$.	Smith and Solly measured (100):$(h0l)$.	Calculated (100):$(h0l)$.	True cell ($a/c = 0{\cdot}701$) $(h0l)$.	Pseudo-cell ($a/c = 4{\cdot}672$) $(h0l)$.
$(50\bar{1})$	11° 39′	11° 20′	$(70\bar{2})$	—
(301)	15° 34′	15° 39′	(502)	—
$(70\bar{2})$	15° 47′	15° 39′	$(50\bar{2})$	—
$(30\bar{1})$	18° 35′	18° 24′	$(21.0.\bar{1}0)$	$(14.0.\bar{1})$
(705)	29° 13′	29° 17′	(504)	—
$(50\bar{3})$	32° 31′	32° 30′	$(11.0.\bar{1}0)$	—
$(40\bar{3})$	42° 25′	43° 3′	$(30\bar{4})$	$(50\bar{1})$
(405)	42° 41′	43° 3′	(304)	(501)
$(60\bar{5})$	44° 4′	43° 56′	$(8.0.\bar{1}1)$	—
(305)	46° 23′	46° 27′	(203)	—
(102)	49° 48′	49° 26′	(305)	(401)
$(10\bar{1})$	50° 48′	50° 48′	$(40\bar{7})$	—
(207)	59° 10′	60° 17′	(205)	(803)
(104)	61° 52′	61° 51′	(308)	(502)
(105)	63° 38′	63° 28′	—	(703)
(108)	67° 14′	66° 50′	—	(201)
$(50\bar{9})$	71° 37′	71° 35′	—	$(14.0.\bar{9})$
(001)	74° 8′	74° 4′	(105)	(403)
$(10\bar{2})$	75° 25′	75° 5′	—	$(50\bar{4})$
(409)	78° 1′	77° 58′	—	(101)
$(30\bar{7})$	79° 37′	79° 41′	—	$(60\bar{7})$
$(20\bar{5})$	80° 47′	80° 53′	—	$(30\bar{4})$
$(10\bar{4})$	85° 10′	85° 6′	—	$(20\bar{5})$
$(10\bar{4})$	86° 58′	86° 56′	—	$(10\bar{3})$
$(3.0.\bar{1}0)$	87° 32′	87° 33′	—	$(10\bar{5})$

A question of the validity of the so-called 'law of rational indices' does not arise. 'Rational' indices can always be found: it is merely a question of the discrepancies between measured and calculated angles that are allowable and the simplicity of indices demanded by the prejudice of the crystallographer. In the present case, only 14 forms of a total of 102 receive indices higher than 11, and of those that receive higher indices 8 are described as bad or striated, so that they should not cause concern.

Fig. 1 shows the gnomonic projection of Smith and Solly's crystal no. 1 on (010). Faces which receive simpler indices on the elements of the pseudo-lattice are limited to certain parts of the projection. Although two lattices are needed for a satisfactory indexing of the forms of sartorite, they are not incongruent lattices. In a discussion of the structure

FIG. 1. Gnomonic projection of sartorite on (010). Full symbols designate forms indexed with reference to the true cell and open symbols designate forms indexed with reference to the pseudo-cell.

and morphology of calaverite, it has been pointed out by Tunell and Ksanda[1] that only some of the faces of that mineral can be simply indexed on the elements of the unit cell and that the other faces are related to 'adventive' spots found in the X-ray patterns. In the case

TABLE III. Indexing of the faces in the [001] zone of sartorite.

($hk0$).	Smith and Solly measured (100):($hk0$).	Calculated (100):($hk0$).	True cell ($a/b = 7{\cdot}494$) ($hk0$).	Pseudo-cell ($a/b = 2{\cdot}498$) ($hk0$).
(11.1.0)	12° 54′	12° 48′	(33.1.0)	(11.1.0)
(710)	19° 35′	19° 38′	(21.1.0)	(710)
(510)	26° 39′	26° 33′	(15.1.0)	(510)
(310)	39° 34′	39° 47′	(910)	(310)
(520)	45° 0′	44° 59′	(15.2.0)	(520)
(210)	51° 8′	51° 19′	(610)	(210)
(110)	68° 1′	68° 11′	(310)	(110)
(120)	78° 24′	78° 41′	(320)	(120)
(010)	89° 41′	90° 00′	(010)	(010)

of sartorite, it may be said that many faces may be simply indexed on the elements of the small pseudo-cell, while others receive simpler indices when referred to the large true cell and are thus related to the spots of the X-ray patterns which indicate the existence of this larger true cell. In a recent paper on meneghinite, a sulpho-salt with one large cell dimension and an unusual ratio, Palache[2] refers to the occurrence of 'aberrant' forms possibly analogous in some measure to the forms that have caused trouble in dealing with sartorite. This may be an indication that such forms are attendant upon certain relations of the cell dimensions.

A recalculation of Trechmann's crystal no. 1 from the published measured angles showed that it may be completely indexed with reference to the structural elements found for sartorite, again using elements corresponding to either the true or pseudo-cell. The crystallographic calculations check the conclusion from the X-ray observations that sartorite-α is identical with sartorite and a comparison of the form lists for the two shows the reason for the previous uncertainty. Of twenty-five forms found in the orthodome zone [010] of sartorite, only four occur on sartorite-α. The latter, however, possesses thirteen forms not found on other sartorite crystals. This shows in a striking fashion that the elements of Trechmann and of Smith and Solly differed as they did because they were designed to fit the peculiarities of particular crystals.

[1] G. Tunell and C. J. Ksanda, Journ. Washington Acad. Sci., 1936, vol. 26, p. 509. [M.A. 7–81.]

[2] C. Palache, Amer. Min., 1938, vol. 23, p. 823.

The relation of stellerite and epidesmine to stilbite.[1]

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[Read November 3, 1938; communicated by F. A. Bannister.]

THE recent finding of a stellerite-like mineral in California has led to this study of its relations to epidesmine and stilbite. The mineral as found by Mr. Allen Nichol, mineralogist of the Division of Highways, the quarry of the Calavera Materials Company near Oceanside, San ego County. Some specimens reached the writer through the State vision of Mines. The mineral occurs as sheaf-like clusters of small stening white crystals in cracks and cavities of andesite. The largest stals are nearly a quarter of an inch long, but tend to aggregate mation so that they do not lend themselves to accurate goniometric easurement. They are apparently orthorhombic, of tabular habit, owing the three pinakoids and small bright faces of a pyramid. The cellent pinakoid cleavage has a pearly lustre, and if this is taken as 10) the crystals, which show parallel extinction, have the same optical entation as stellerite.

Stellerite was described from the Commander Islands, Bering Sea, by orozewicz[2] and classed as orthorhombic at least in part because of its rallel extinction. Measurement of the rather poor crystals led to an ial ratio similar to the one given for stilbite by Victor Goldschmidt d others using the orthorhombic setting of that mineral. This ilarity seems to have been overlooked by Morozewicz, who rightly sidered stilbite to be monoclinic. The formula he assigned to steller-, $CaAl_2Si_7O_{18}.7H_2O$, differed from the formulae then usually given stilbite, but it proves to be just a special case of the general stilbite mula which will be referred to again below.

Stellerite has since been reported from near Juneau, Alaska, by heeler,[3] who obtained the same axial ratio as Morozewicz, but also

[1] Work done in the Mineral Department of the British Museum during leave of ence as Fellow of the John Simon Guggenheim Memorial Foundation.

[2] J. Morozewicz, Über Stellerit, ein neues Zeolithmineral. Bull. Intern. Acad. Cracovie, Cl. Sci. Math. Nat., 1909, 2me semestre, pp. 344–359.

[3] E. P. Wheeler, Stellerite from near Juneau, Alaska. Amer. Min., 1927, vol. 12, 360–364. [M.A. **4**–325.]

failed to notice its similarity to that of stilbite in the orthorhombic setting.

Epidesmine was described from Schwarzenberg, Saxony, by Rosický and Thugutt[1] as an orthorhombic mineral having the same formula '$3Ca(Na_2,K_2)Al_2Si_6O_{16}.20H_2O$', as stilbite (German—Desmin). The crystals showed mainly just three pinakoids at right angles and the parallel extinction which seemed to confirm the orthorhombic character.

Epidesmine has since been recorded from various localities by several American mineralogists. Shannon,[2] describing 'stilbite of "epidesmine" habit bounded only by three pinacoids', affirmed the identity of stilbite with epidesmine, but, following Goldschmidt, considered them to be orthorhombic. He did point out that the optical orientation of epidesmine becomes similar to that of stilbite if the most prominent cleavage is chosen as (010), rather than (100), as was done by Rosický and Thugutt. It may be noted that this also makes the optical orientation identical with that of stellerite as given by both Morozewicz and Wheeler.

Stilbite has generally been considered to be monoclinic, but a number of mineralogists, including Victor Goldschmidt,[3] have treated it as orthorhombic because of its peculiar habit. A recent X-ray study of stilbite[4] has shown definitely that it is monoclinic and has led to an improved formula based on the count of the cell contents of many samples.

It seems reasonable to conclude that stellerite and epidesmine are identical with stilbite and are merely pseudo-orthorhombic. In the following it will be shown that this conclusion is in harmony, not only with the X-ray evidence, but also with the goniometric, chemical, and optical evidence, if the difficulties of good interfacial angle measurements and the variability of composition and properties of such material as stilbite be taken into account.

In the monoclinic setting stilbite is given A. von Lasaulx's (1878) elements[5] $0{\cdot}7623:1:1{\cdot}1940$, $\beta = 50° 50'$. It is commonly elongated parallel to the a-axis. If set up with this axis vertical the three forms

[1] V. Rosický and S. J. Thugutt, Epidesmin, ein neuer Zeolith. Centralbl. Min., 1913, pp. 422–426.

[2] E. V. Shannon, The mineralogy and petrology of intrusive Triassic diabase at Goose Creek, Loudoun County, Virginia. Proc. U.S. Nat. Mus., 1924, vol. 66, art. 2, p. 72. [M.A. **3**–204.]

[3] V. Goldschmidt, Krystallographische Winkeltabellen, 1897, p. 113.

[4] J. Sekanina and J. Wyart, Sur la stilbite. Bull. Soc. Franç. Min., 1936, vol. 59, pp. 377–383. [M.A. **6**–525.]

[5] As quoted by Dana, System of mineralogy, 6th ed., 1892, p. 583, and in Dana Textbook of mineralogy, 4th ed., 1932, p. 648.

01), (010), and ($\bar{1}$01), will simulate three orthorhombic pinakoids since ie interfacial angles are: (001):(010) = 90° 0′, ($\bar{1}$01):(010) = 90° 0′, 01):(001) = 89° 30′. The forms ($\bar{1}$12) and ($\bar{1}$10) together will then mulate an orthorhombic pyramid as shown in the stereographic pro- ction fig. 1. The faces of these two forms were taken by Morozewicz id by Wheeler to belong to the orthorhombic pyramid (111).

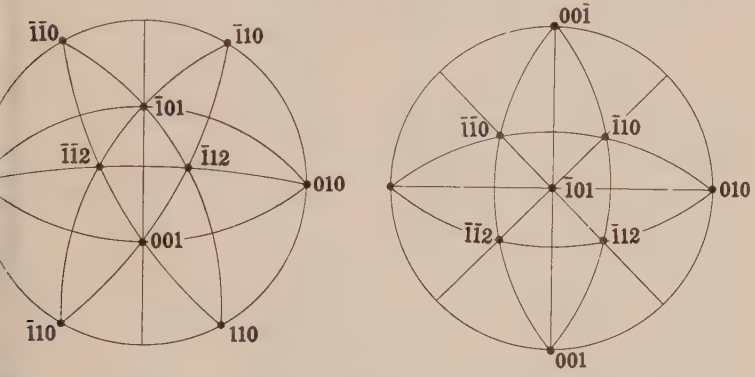

Fig. 1. Stereographic projections of stilbite on a plane normal to the c-axis and a plane normal to the a-axis, the latter showing the pseudo-orthorhombic habit. e projection is brought into the position of the other by a rotation of 50° 50′ about b-axis.

The angles calculated from A. von Lasaulx's elements are:

($\bar{1}$01):($\bar{1}$12) = 48° 2′	(001):($\bar{1}$12) = 56° 28′	(010):($\bar{1}$12) = 59° 45′
($\bar{1}$01):($\bar{1}$10) = 48 30	(001):($\bar{1}$10) = 57 3	(010):($\bar{1}$10) = 58 25

the similar angles are averaged, as if the mineral were considered horhombic, and the (orthorhombic) axial ratio calculated from the raged angles, the result with the pyramid as (111) is $a:b:c$ = 23:1:0·761. This is close to the values, 0·928:1:0·762, given by ldschmidt and differs a little more from the values, 0·98:1:0·76 and 82:1:0·762, given by Morozewicz and Wheeler respectively. The crepancy is easily explained when one considers the character of the stals. Both Wheeler and Morozewicz relied partly on angles measured der the microscope. The writer found that in the measurement of en crystals of the San Diego County material readings on similar erfacial angles varied over a range of three or four degrees.

Table I gives the compositions and densities of the original stellerite d epidesmine, of Shannon's 'stilbite of "epidesmine" habit', and of stellerite-like stilbite from San Diego County. All the densities given within the limits, 2·094–2·205, recorded by Dana for the density of

stilbite. The compositions all agree with the formula, $Na_xCa_{2+y}Al_{4+x+2y}Si_{14-x-2y}O_{36}.14H_2O$, with $x < 1$ and $-0{\cdot}15 < y < 0{\cdot}30$, found by Sekanina and Wyart for stilbite from the examination of the contents of the unit cell in many samples.

TABLE I. Chemical analyses.

	I Stellerite.	II Epidesmine.	III 'Stilbite of "epidesmine" habit'.	IV Stellerite-like stilbite.
SiO_2	59·23	56·66	54·40	58·48
Al_2O_3	14·41	16·00	17·88	14·76
Fe_2O_3	0·22	—	—	0·48
MgO	—	0·06	—	0·32
CaO	8·23	7·58	8·56	7·68
K_2O	—	0·67	—	0·29
Na_2O	trace	0·88	—	1·02
H_2O+	18·15	18·69	16·00	12·30
H_2O-			2·32	4·43
Insoluble	—	0·44	—	--
Total	100·24	100·98	99·16	99·76
Sp. gr.	2·124	2·152	—	2·120

I. Copper Island, Commander Islands, Bering Sea. J. Morozewicz (1909).
II. Gelbe Birke mine, Schwarzenberg, Saxony. V. Rosický and S. J. Thugutt (1913).
III. Goose Creek, Loudoun County, Virginia. E. V. Shannon (1924).
IV. Oceanside, San Diego County, California. W. H. Herdsman, analyst.

In table II are shown the proportions of the constituents in the four samples here considered on the basis of oxygen equal to 36. It will be seen that the values of x and y are in all cases well within the given limits, and that the sum of Si and Al is always quite close to 18. The variation in the water content should not be disturbing, since Sekanina and Wyart found the number of water molecules in stilbite to vary from 13·35 to 14·9.

TABLE II. Contents of unit cell on the basis of 36 atoms of oxygen.

	(Na,K).	Ca.	Al.	Si.	O.	H_2O.	x.	y.	Si+Al
I.	0·00	2·08	4·00	13·96	36	14·24	0·00	+0·08	17·96
II.	0·61	1·96	4·54	13·45	36	15·00	0·61	−0·04	17·99
III.	—	2·20	5·07	13·10	36	14·70	—	+0·20	18·17
IV.	0·56	1·95	4·12	13·76	36	13·26	0·56	−0·05	17·88

Table III shows the optical properties of the specimens whose analyses are considered above together with those of stilbite. The orientation is the same in every case, the optic axial plane being parallel to the bes

avage and α sensibly parallel to the elongation, except for stilbite ich is usually stated to have $\alpha:c = 5°$.

Stellerite and epidesmine were originally taken to be orthorhombic rtly on account of their parallel extinction. It may now be said that e small extinction-angle of stilbite varies with differences in com- sition and that stellerite and epidesmine are varieties in which this gle is sensibly zero.

TABLE III. Optical data.

	I.	II.	III.	IV.	Stilbite.
α ...	1·484		1·490	1·486	1·494[2]
β ...	1·49	mean 1·498[1]	1·500	1·494	1·498
γ ...	1·495		1·502	1·496	1·500
Sign ...	negative	negative	negative	negative	negative
2V ...	43½°	moderate	medium	48°	33°±
bserver ...	Morozewicz	Rosický and Thugutt	Shannon	Pabst	—

[1] E. S. Larsen and H. Berman, The microscopic determination of the nonopaque nerals, Bull. U.S. Geol. Surv., 1934, no. 848, give the values 1·493, 1·501, and 07 for epidesmine, and other observers have given slightly different values.
[2] As given by Larsen and Berman, ibid.

The writer has been fortunate in being able to obtain samples of the ginal stellerite from the Commander Islands and of the original desmine from Schwarzenberg, Saxony, in the mineral collection of British Museum for comparative X-ray study.

Rotation patterns on the axis of elongation of these two materials and the stilbite from San Diego County proved to be identical with rota- n patterns of excellent crystals of stilbite from the well-known locality Naalsö, Faeroe Islands. Dimensions calculated from these patterns ee with the unit cell given by Sekanina and Wyart, a 13·60, b 18·13, 1·29 Å., β 52°.

Laue patterns taken normal to (010) of the original stellerite are like ose taken normal to (010) of twinned stilbite. As Sekanina and Wyart inted out, stilbite may be assigned a pseudo-orthorhombic cell with nearly 90°. This corresponds to the pseudo-orthorhombic description ferred to above. The patterns of twinned stilbite and of stellerite pear orthorhombic. Lack of doubling of spots in the [100] and [101] nes in these patterns shows that the angle β of the pseudo-orthorhombic l is sensibly 90°. Other patterns of untwinned stilbite, obtained only er repeated trials in an effort to restrict the passage of the incident am to a single untwinned part of a crystal, show that stilbite is truly noclinic and that the orthorhombic appearance of the other patterns

is merely due to the repetition of spots by twinning. Unfortunately the crystals of epidesmine, though yielding an excellent rotation pattern whereby their relation to stilbite could be established, appear to have such a bundle structure that no good Laue patterns are obtainable from them.

The writer is very grateful to Mr. W. Campbell Smith, Keeper of Minerals, for his hospitality and for permission to use the facilities of the Department of Mineralogy of the British Museum. He is greatly indebted to Mr. F. A. Bannister for the preparation of X-ray patterns and for a critical reading of the manuscript. The cost of analysis of the stilbite from San Diego County was defrayed by a grant from the Board of Research of the University of California.

Sapphirine crystals from Blinkwater, Transvaal.

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[Read January 26, 1939.]

SOME time ago, Mr. F. C. Partridge, Mineralogist to the Geological Survey of South Africa, gave me some crystals of sapphirine which thought might be of interest for goniometric examination. They me from the farm Blinkwater 1037, about 47 miles NNW. of Pietersrg and 45 miles WSW. of Louis Trichardt in northern Transvaal, and re collected by him in July 1930, after he had identified some specins of sapphirine sent to him by Mr. A. Ruskovitch of Abrasive, thern Transvaal. They occur in a coarse granitic rock rich in brown a, and are associated with small prismatic crystals of pink corundum. The sapphirine crystals are almost invariably tabular or platy in bit, although occasional crystals possess a rhomb-shaped cross-ion. The largest crystal measures $50 \times 35 \times 5$ mm., while the thickest is only 7 mm. in thickness. They are generally six-sided in outline th a strong suggestion of orthorhombic symmetry.

Chemical tests on fused material showed the crystals to consist sentially of silica, alumina, and magnesia. The average specific gravity several isolated crystals was determined as 3·35, and the hardness on freshly fractured surface as $6\frac{1}{2}$. As the material, however, showed siderable alteration when examined optically, neither of these figures uld be regarded as characteristic. The crystals are normally dull in re and of a pale grey colour or various shades of green, red, and rple on the surface. In addition, they sometimes carry tiny mica kes, in various stages of alteration, adhering parallel to the large es. A few crystals, however, possess fairly bright faces which then pear greyish-black with a bluish tinge. Even these faces rarely give rp images, so that interfacial angles can generally be determined only the nearest whole degree; and where the angles can be more accuely determined corresponding readings show discrepancies considerly greater than half a degree. In some cases where faces were smooth t dull, attempts were made to determine interfacial angles by attach- tiny pieces of cover-glass to them, but the results again showed but

little agreement. The results, then, can be relied upon only to the nearest degree.

The crystals are really monoclinic and the common habit, shown in fig. 1, consists of the following forms: *b* (010), *u* ($\bar{1}32$), *m* (110), *e* (102), *q* (011), *f* (012), *a* (100), *c* (001), and *d* ($\bar{1}02$), while tiny faces were observed parallel to *k* (270), *s* ($\bar{1}12$), *t* ($\bar{1}22$), and *v* ($\bar{1}42$). In lettering the faces, I have adopted the suggestions of T. V. Barker ('Graphical and tabular methods in crystallography', 1922, p. 97). Two less common habits are shown in figs. 3 and 4, projected on the clinopinakoid face. Nearly all the crystals show prominent striations, ridges, and furrows, due to parallel growth, on the clinopinakoid faces parallel to the [201] zone. The crystals show two very imperfect cleavages parallel to *a* (100) and *c* (001), which are, however, not recognizable in thin section.

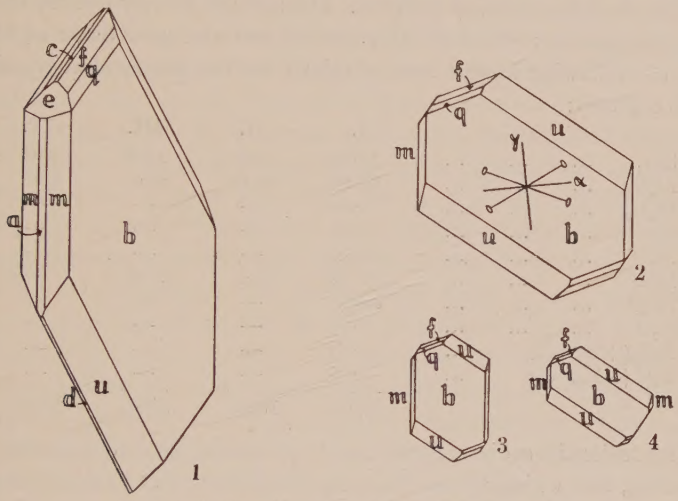
The values obtained for the interfacial angles are given in the first column of the following table:

	E. D. Mountain (1938) measured.	A. Lacroix (1913) measured.	A. Lacroix (1913) calculated.
ae = (100):(102)	34°	34° 35′	33° 53′
ec = (102):(001)	34	—	34 40
cd = (001):($\bar{1}02$)	55	—	56 33
bk = (010):(270)	25	—	24 1
bm = (010):(110)	57	57 20	57 20
ba = (010):(100)	90	—	90 0
bq = (010):(011)	38	37 2	37 24
bf = (010):(012)	58	57 5	56 49
bc = (010):(001)	90	—	90 0
bv = (010):($\bar{1}42$)	24	—	23 31
bu = (010):($\bar{1}32$)	31	—	30 7
bt = (010):($\bar{1}22$)	42	—	41 2
bs = (010):($\bar{1}12$)	61	—	60 7
bd = (010):($\bar{1}02$)	90	—	90 0

Owing to the similarity in angles and development between the prism and clinodome zones, it was felt that there was in the case of most crystals insufficient evidence to discriminate definitely between the two zones on goniometric data alone. For this reason, several sections were cut parallel to the clinopinakoid in order to orientate the crystals according to their optical properties. The material is never perfectly fresh and is generally altered to a yellowish decomposition product along a network of cracks, so that only thin sections up to second- or third-order interference-colours were sufficiently transparent for optical determination. The optical orientation is shown in fig. 2.

A section, parallel to the optic axial plane, which gave first-order

sensitive tint when examined in convergent light, showed that the α vibration-direction corresponds to the acute bisectrix, thus demonstrating its optically negative character.[1] These sections parallel to the clinopinakoid show very conspicuous dichroism in blue and buff.



FIGS. 1–4. Sapphirine crystals from Transvaal.

Another section was prepared parallel to the orthopinakoid *a* (100) and when examined with a high-power objective exhibited both optic axes within the field but somewhat unequally inclined to the normal. This section was used to determine the optic axial angle both by the Mallard method and on the universal stage, the two results agreeing perfectly. In this section dichroism was only slight.

The optical characters determined on the same crystal may be summarized as follows:

γ to vertical axis *c* [001] = 6° in the obtuse axial angle; $\beta = b$ [010].

α 1·714, β 1·719, γ 1·720 (by immersion).

α pale pinkish-buff, β pale cerulean-blue, γ Italian-blue.

$2V_\alpha = 50\frac{1}{2}°$, negative, $r < v$, inclined dispersion.

The optical characters enabled the crystallographic orientation to be determined in all cases, and the following general distinctions were then drawn between the prism and clinodome zones:

(1) *a* (100) well developed, *c* (001) poor.

(2) re-entrant angles rare in prism zone, common in clinodome zone.

(3) rounded faces rare in prism zone, common in clinodome zone.

[1] E. D. Mountain, South African Journ. Sci., 1929, vol. 26, p. 70. [M.A. 4–443.]

(4) $bk = 25°$, while $bq = 38°$.

(5) faces in the prism zone are generally better developed than in the clinodome zone.

The crystal which had been used for the optical properties was then further used for chemical analysis. The specific gravity determined by the pyknometer method on $3\frac{1}{2}$ grams of material gave 3·398 at 11° C. and the following figures were obtained for the chemical composition (first column):

	I.	II.	III.	IV.
SiO_2	12·95	13·9	12·9	13·1
Al_2O_3	62·38	67·0	65·7	66·5
Fe_2O_3	1·69	—	—	—
FeO	3·09	—	—	—
MnO	trace	—	—	—
MgO	15·22	19·1	21·4	20·4
Na_2O	trace	—	—	—
K_2O	0·10	—	—	—
H_2O (+110° C.) ...	4·80	—	—	—
H_2O (−110° C.) ...	0·05	—	—	—
	100·28	100·0	100·0	100·0

The fusion in sodium carbonate took several hours even after fine grinding, and owing to the very slight solubility the value for ferrous oxide may not be quite reliable. Another crystal which possessed a specific gravity of 3·146 was found to contain 10·41% combined water. In the second column the analysis has been recalculated to a water-free basis, the iron oxides being represented by an equivalent amount of MgO.

For comparison, in the third column of the above table, are given the figures corresponding to $5MgO.6Al_2O_3.2SiO_2$, the generally accepted formula for sapphirine. Gossner and Mussgnug[1] have shown that the crystal-structure of sapphirine corresponds to the formula $Mg_2Al_4SiO_{10}$ in which MgSi is further indefinitely replaced by Al_2. Actually, this agrees more closely with our recalculated composition when, in $4Mg_2Al_4SiO_{10}$, just one MgSi is replaced by Al_2. This would actually be equivalent to $Mg_7Al_{18}Si_3O_{40}$, and the percentage composition corresponding to it is given in column IV.

Data published in the literature on sapphirine show considerable differences. Some of the original articles were inaccessible to me and I am greatly indebted to Dr. L. J. Spencer for a number of critical summaries. The earliest crystallographic data were obtained on crystals

[1] B. Gossner and F. Mussgnug, Neues Jahrb. Min., Abt. A, 1928, Beil.-Bd. 58, pp. 233–238. [M.A. 4–109.]

m Fiskernæs in Greenland by Ussing,[1] who gives an average value
bm of 57° 27′ and an extinction-angle $\gamma:c$ 8½° in the acute axial angle.
is, together with his axial angle of 79½° and an angle of 47½° for
0):(011), suggests that he adopted a different clinoaxis from ours, but
e same prism zone. Unfortunately, the axial angle was only deduced
m measurements made on different crystals and not determined
ectly. To add to the confusion, later authors misquote Ussing's data.
r instance Dana[2] is not only himself inconsistent but implies an
inction-angle of 19° in the obtuse axial angle. Walker and Collins[3]
1907 described crystals of sapphirine from Vizagapatam, Madras,
th an extinction-angle of about 15°, but the orientation is not clear.
In 1913 A. Lacroix[4] described in some detail crystals of sapphirine
m near Itrongay in Madagascar. They showed the forms *b a e m f* and
and his crystallographic calculations were based on the measured
gles *am* 32° 40′, *ef* 46° 30′, *fm′* 92° 10′. From these values he deduced
$b:c = 0{\cdot}7310:1:1{\cdot}4115$, $\beta = 68°\ 33'$, but although the axial angle is
rectly calculated the axial ratios should read $a:b:c = 0{\cdot}6889:1:1{\cdot}4056$.
ave employed these latter figures, and the calculated angles given in
table above have been based on them. The agreement between these
culated angles and my measured angles is reasonably good, so that
an be assumed that the elements upon which the calculations were
ed are fairly accurate. There seems to be no reason why the length
he vertical axis should be more than twice that of the clinoaxis, and
he vertical axis were halved the indices would actually be simplified;
to avoid any confusion I have refrained from introducing any such
nge. Lacroix did not determine the optic orientation but adopted
sing's 8½° and assumed that Ussing's (001) corresponded to our ($\bar{1}04$).
th our orientation, this would correspond to an angle for $\gamma:c$ of 8½°
8° after Lacroix) in the obtuse, instead of the acute, axial angle.
roix expresses this by saying that the negative acute bisectrix
kes an angle of 98° with the vertical axis in the acute angle ph^1, i.e.
4):(100), a statement quite liable to misconstruction.

A. N. Winchell[5] has interpreted the foregoing results for the extinction-
le $\gamma:c$ as being +8° (Lacroix) and 15° (Walker), thus in the obtuse

[1] N. V. Ussing, Öfversigt K. Vetenskaps-Akad. Förhandl., 1889, vol. 46, pp. 17–
modified translation in Zeits. Kryst. Min., 1889, vol. 15, pp. 598–605, 2 figs.
[2] E. S. Dana, System of mineralogy, 6th edit., 1892, p. 561.
[3] T. L. Walker and W. H. Collins, Rec. Geol. Survey India, 1907, vol. 30, p. 8.
[4] A. Lacroix, Min. de France, 1913, vol. 5, pp. 73–74, 1 fig.
[5] A. N. Winchell, Elements of optical min., 2nd edit., 1927, pt. 2, p. 277; 3rd edit.,
3, pt. 2, p. 427.

axial angle. In the text he gives $\beta = 68° 33'$ following Lacroix, but his figure is drawn with $\beta = 79\frac{1}{2}°$ as given by Ussing. On the other hand Larsen[1] gives $\gamma:c = -8\frac{1}{2}°$ (i.e. in the acute axial angle), apparently following Ussing and ignoring later authors.

More recently, Gossner and Mussgnug (loc. cit.) attempted to reconcile some of these discrepancies. They remeasured Ussing's original crystals but, as before, the prism zone and the clinodome zone were measured on different crystals. They calculated β as 74° 31′ and $a:b:c =$ 0·667:1:0·691 based on X-ray measurements. They have halved the vertical axis, as suggested above, but their axial angle receives no support from the crystals described in this paper.

In conclusion, it is suggested that Lacroix's measurements be accepted as substantially correct giving $a:b:c = 0{\cdot}6889:1:1{\cdot}4056$, $\beta = 68° 33'$ while the optic orientation be defined as optic axial plane parallel to b (010) with $\gamma:c = 6°$ in the obtuse axial angle.

[1] E. S. Larsen, Bull. U.S. Geol. Surv., 1934, no. 848.